Basic Aquifer parameters

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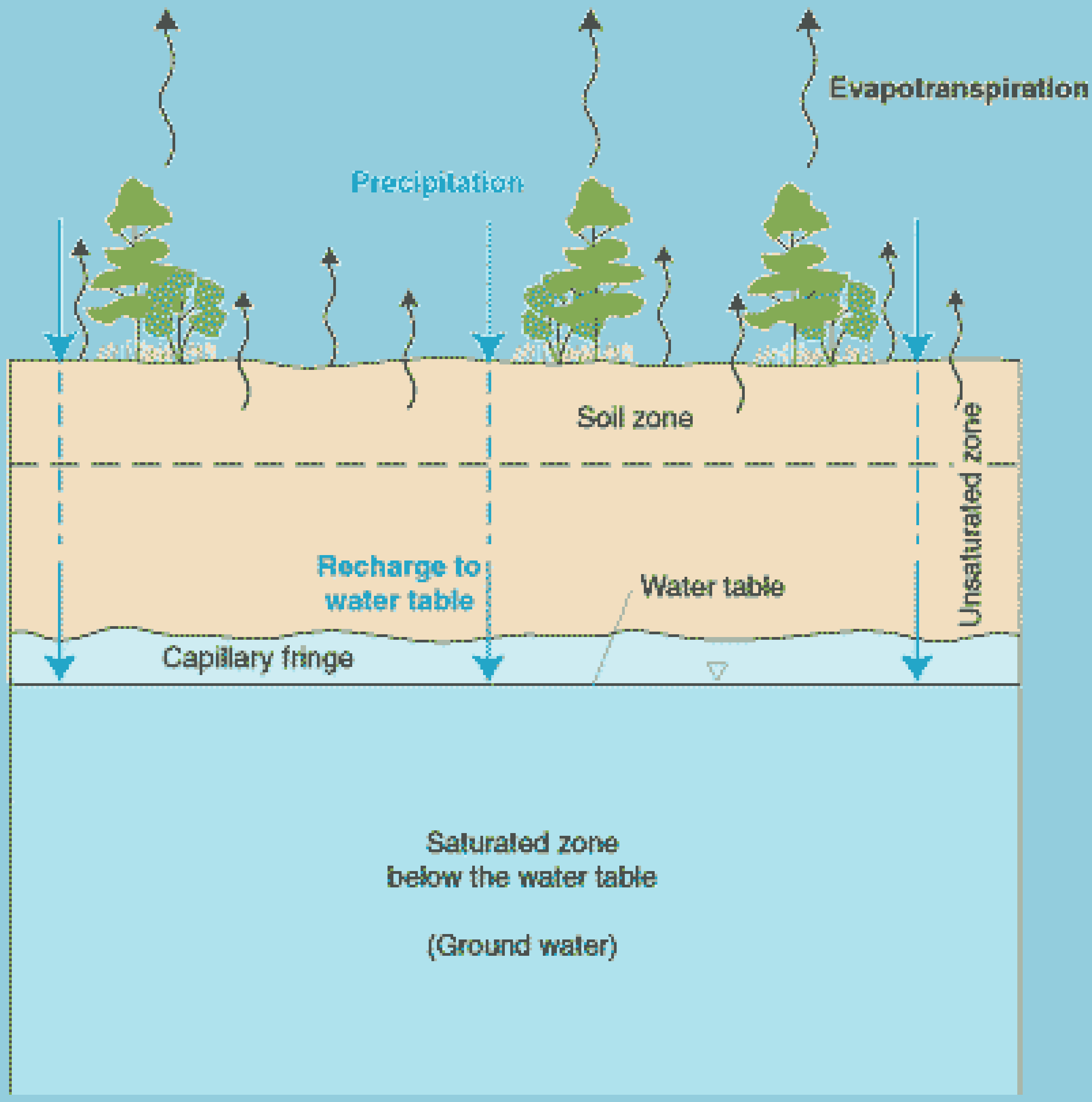
Underground water

- "Underground water occurs in two different zones. One zone, which occurs immediately below the land surface in most areas, contains both water and air and is referred to as the unsaturated zone.
- It is the zone between the land surface and the deepest water table. It includes the capillary fringe.
- The unsaturated zone, is also known as the terms 'zone of aeration' and 'vadose zone,'

Unsaturated zone

- The unsaturated zone is frequently divided into three components.
- **Soil Zone**:- The first zone is the soil zone, generally a meter or two thick, which contains living roots and which supports plant growth. The amount of moisture remaining in the soil zone after an extended period of drainage is called ***‘field capacity’***.
- **Intermediate zone**:- It is the second zone not to be confused with the 'intermediate' zones of saturated flow systems.
- **Capillary Zone**:- The boundary between the unsaturated zone and the saturated zone is termed the capillary fringe." - Chapelle, 2001



Evapotranspiration
Precipitation
Soil zone
Unsaturated zone
Recharge to
water table
Water table
Capillary fringe
Saturated zone
below the water table
(Ground water)

Source: ***USGS***

Saturated zone

- Although a considerable amount of water can be present in the unsaturated zone, this water cannot be pumped by wells because capillary forces hold it too tightly.
- In contrast to the unsaturated zone, the voids in the saturated zone are completely filled with water. The approximate upper surface of the saturated zone is referred to as the water table.
- Water in the saturated zone below the water table is referred to as ground water. Below the water table, the water pressure is high enough to allow water to enter a well as the water level in the well is lowered by pumping, thus permitting ground water to be withdrawn for use.

Types of Water Bearing Formations

Types of Water Bearing Formations

AQUIFER: -

- saturated geological formation that is permeable enough to yield significant quantities of water to wells and springs.
- A porous and permeable water bearing formation is called an *aquifer*.
- The terms "*water bearing formation/stratum*" and "*ground water reservoir*" are synonyms for the aquifer.

Types of Water Bearing Formations

AQUICLUDE:-

- **It is a saturated formation through which virtually no water is transmitted.**
- **Aquicludes may have high porosity but relatively have very low permeability and hence do not yield appreciable quantities of water to wells.**
- **A highly porous and an impervious (that does not transmit water at all) geological formation is called an *aquiclude* e.g. clay and shale.**

Types of Water Bearing Formations

AQUITARD:-

- A saturated formation which yields inappreciable quantity of water to wells but through which appreciable leakage of water is possible is known as ‘aquitard’ or ‘semi-confining layer’.
- Aquitards are rock layers that are partly impervious and transmit water at a lower rate than aquifer (e.g.) sandy clay.

Types of Water Bearing Formations

AQUIFUGE :-

- **It is a geological formation which is neither porous nor permeable.**
- **It neither stores nor transmits water**
- **Massive igneous and sedimentary rocks (compact limestone).**

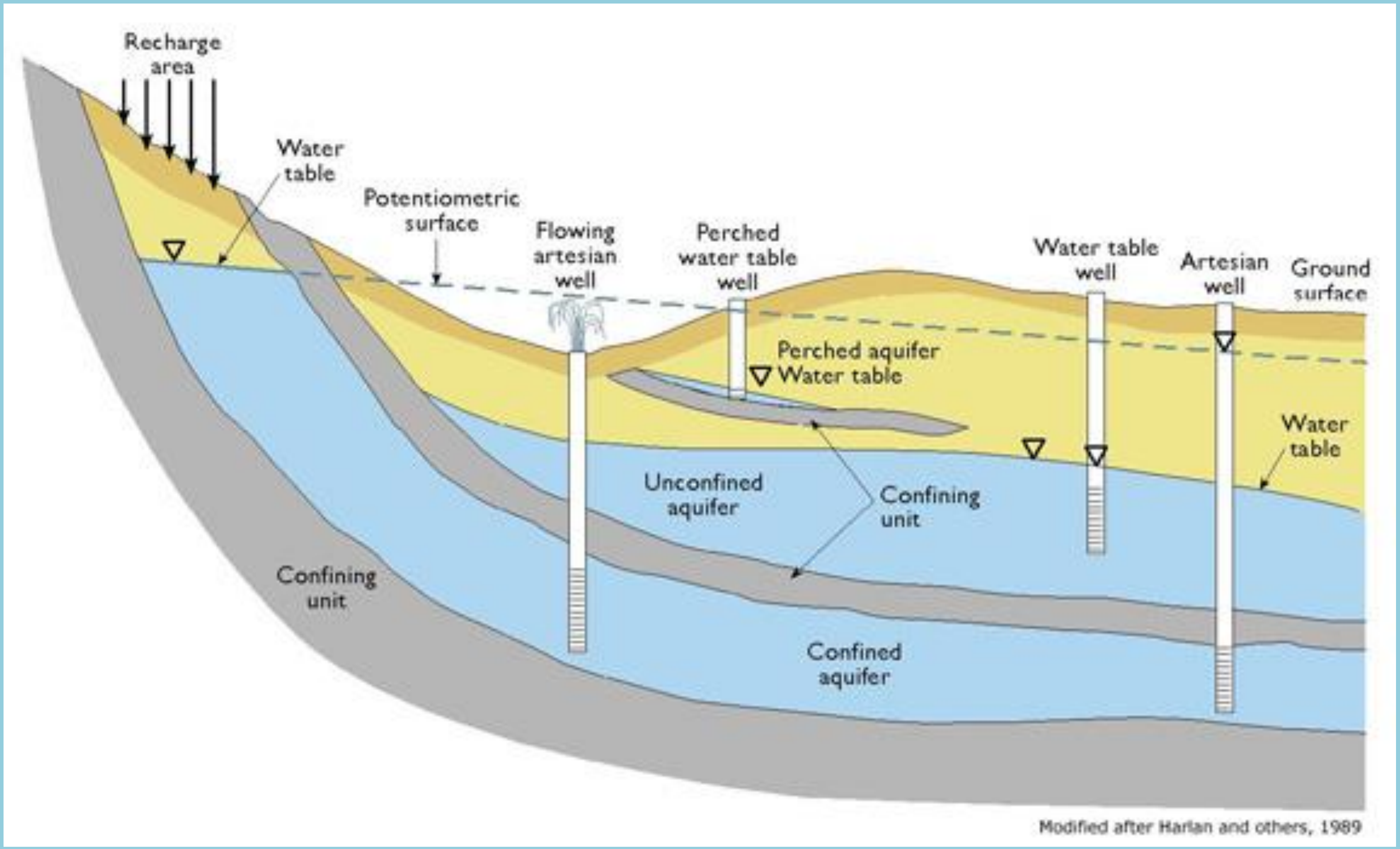
RECAP

AQUIFER	**Saturated** **Porous & Permeable** **Storage capacity** **Water holding and yielding**
AQUICUDE	**Saturated** **Porous & impermeable** **Storage capacity** **Water holding but not yielding**
AQUITARD	**Saturated** **Porous & Semi-permeable** **Storage capacity** **Only seepage is possible**
AQUIFUGE	**Non-saturated** **Non-porous & Non-permeable (impervious/impermeable)** **No storage capacity**

TYPES OF AQUIFERS

TYPES OF AQUIFERS

- **Aquifers in an area can be classified on the basis of their location in the ground water basin, and the position of their associated water levels.**
- **Aquifers are of three types:**
  - **a) Unconfined**
  - **b) Confined**
  - **c) Leaky**

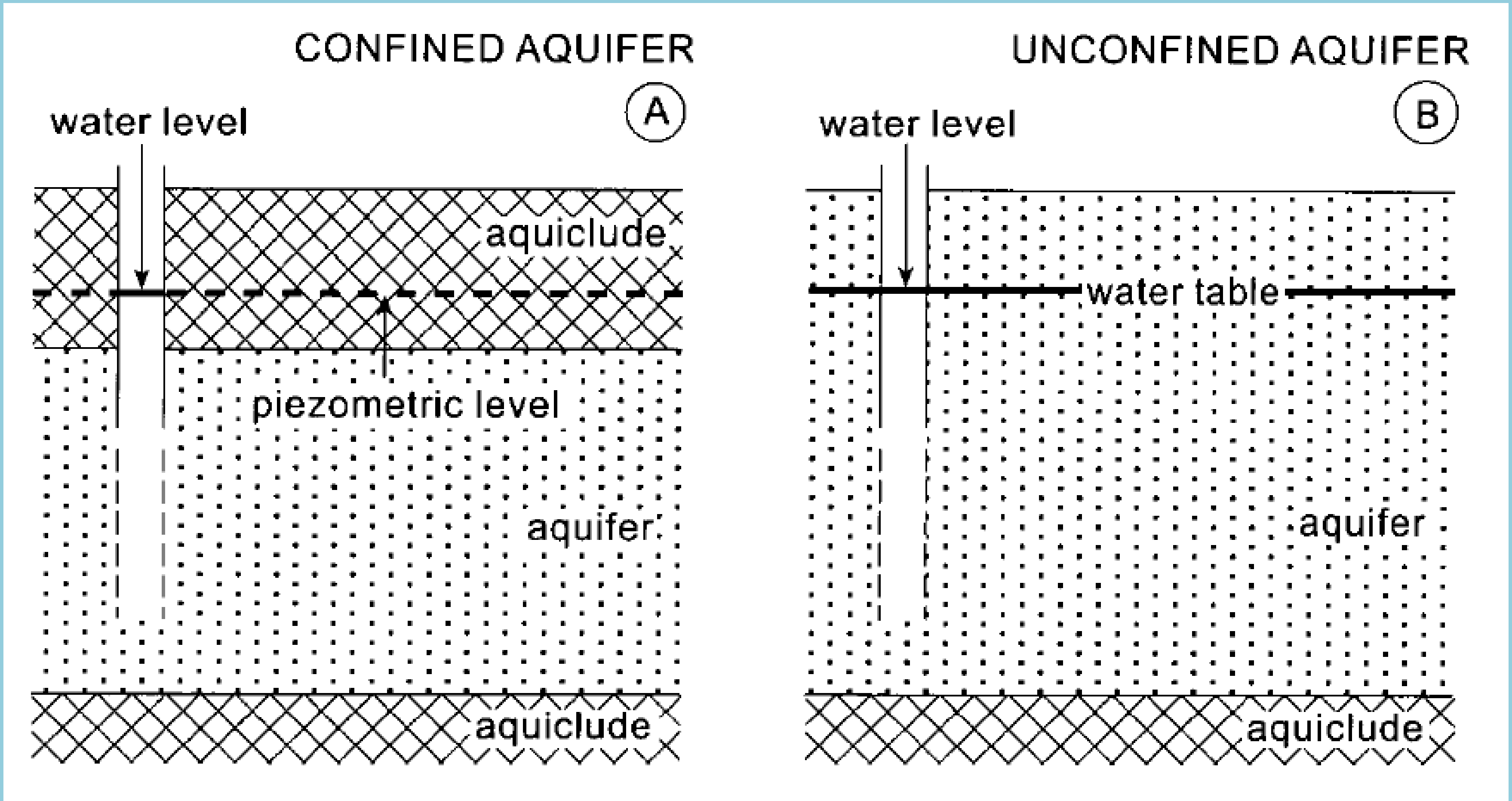

Recharge area
Water table
Potentiometric surface
Flowing artesian well
Perched water table well
Water table well
Artesian well
Ground surface
Perched aquifer
Water table
Water table
Unconfined aquifer
Confining unit
Confining unit
Confined aquifer
Modified after Harlan and others, 1989

UNCONFINED AQUIFER

- **An unconfined aquifer is a partially satuared aquifer. It is not overlain by any confining layer but it has a confining layer at the bottom.**
- **The upper surface is defined by the water table and it is in direct contact with the atmosphere.**
- **The water level in an unconfined aquifer is known as 'phreatic water level'**
- **Also known as *water table* or *phreatic aquifer*.**

UNCONFINED AQUIFER

- **Water in a well penetrating an unconfined aquifer is under atmospheric pressure and therefore does not rise above the water table.**
- **The water table in unconfined aquifers is free to rise and fall.**
- **Rises and falls in the unconfined aquifer correspond to changes in the volume of water in storage within aquifer.**
- **Movement of the ground water is in direct response to gravity.**


CONFINED AQUIFER
A
water level
aquiclude
piezometric level
aquifer
aquiclude
UNCONFINED AQUIFER
B
water level
water table
aquifer
aquiclude

CONFINED AQUIFERS

- A confined aquifer is a completely a saturated aquifer that is bounded above and below by aquicludes, which is impermeable to water flow.
- It has an overlying confining layer.
- The pressure of water in confined aquifers is usually higher than the atmospheric pressure, so that if a well taps the aquifer, the water level will rise above the top of the aquifer i.e. above the base of the overlying confining bed. It will rise up to an elevation at which it is in balance with the atmospheric pressure.

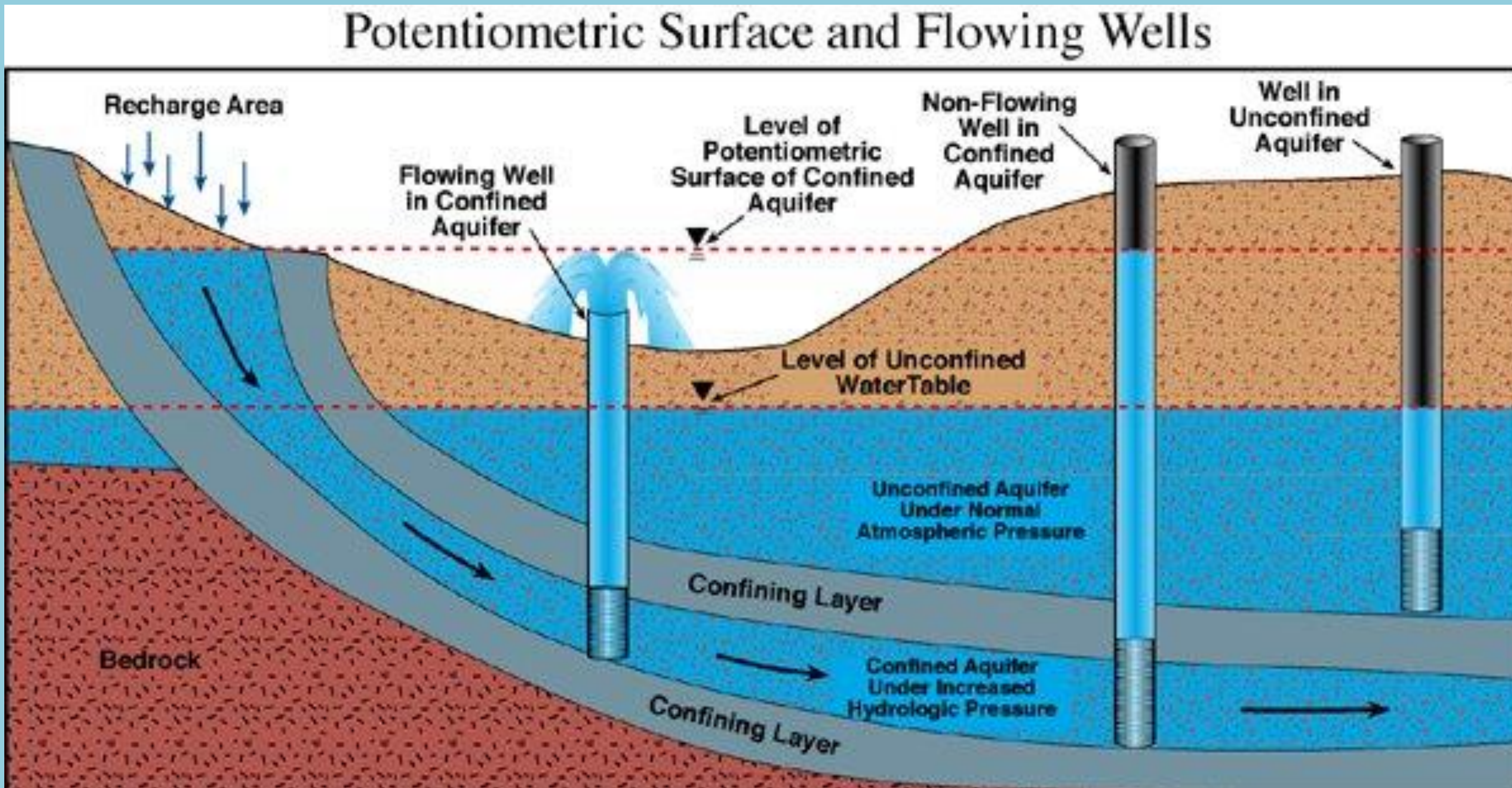
CONFINED AQUIFERS

- **If this elevation is greater than that of the land surface at the well, the water will flow from the well and such wells are termed *artesian* or *flowing wells*.**
- **The imaginary surface, conforming to the elevations to which water will rise in wells penetrating confined aquifers is known as the *piezometric surface or potentiometric surface.* It coincides with the hydrostatic pressure levels of the water in the aquifer.**

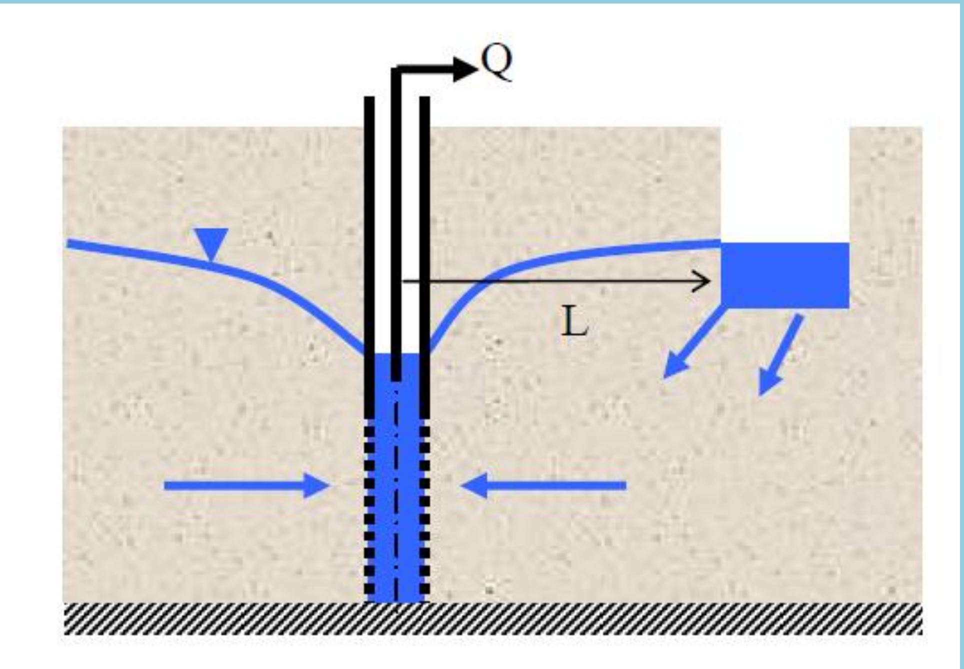
CONFINED AQUIFERS

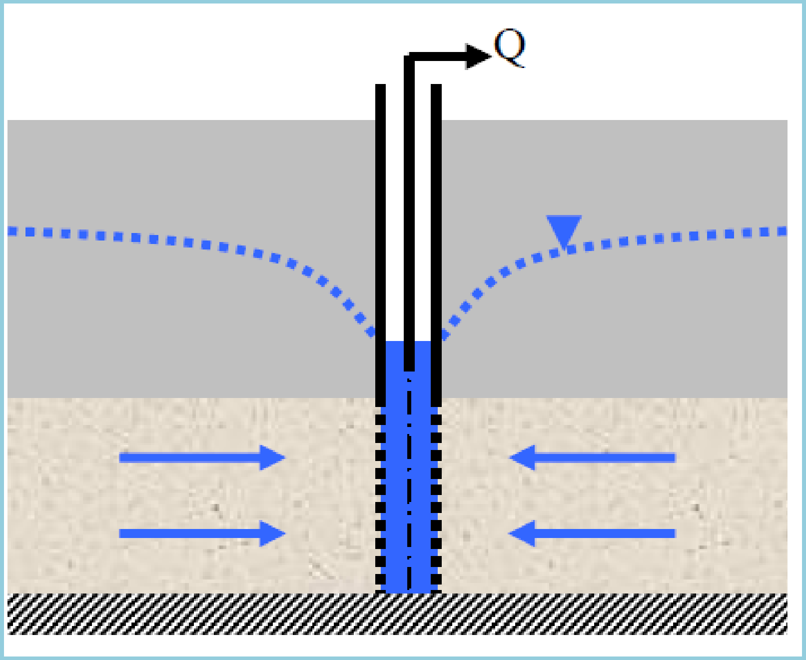
- **Rises and falls of water in wells penetrating confined aquifers result primarily from changes in pressure rather than changes in storage volumes.**
- **Therefore confined aquifers display only small changes in storage.**
- **They mainly serve as conduits for conveying water from recharge to discharge areas.**

CONFINED & ARTESIAN AQUIFERS



Release of water from confined and unconfined aquifers





Release of water from confined and unconfined aquifers (USGS Illustration)

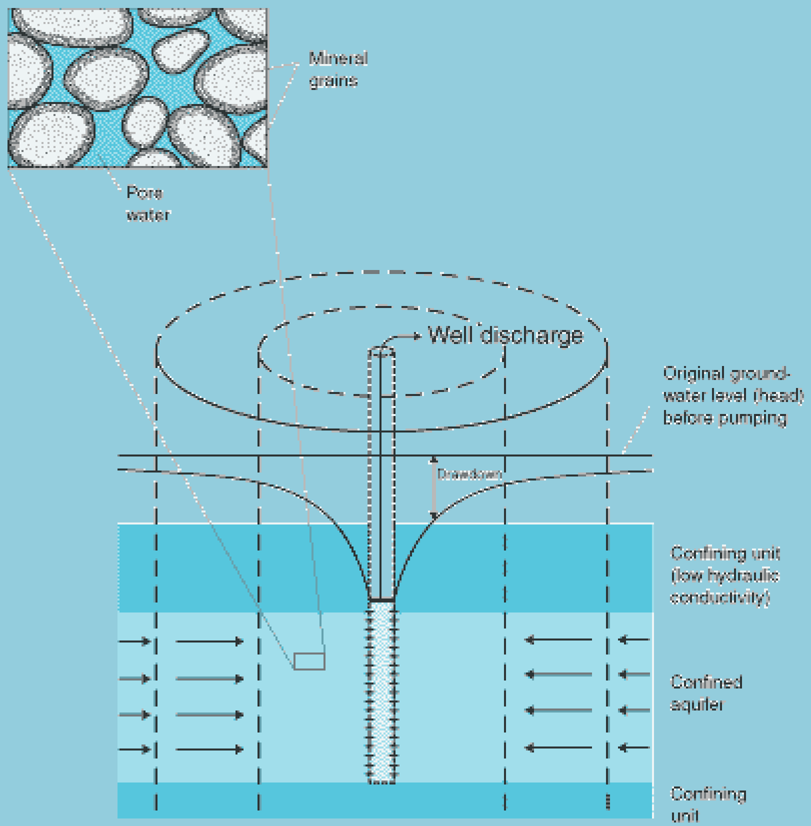


Fig A1:- Pumping a single well in an idealized confined aquifer. Confined aquifers remain completely saturated during pumping by wells (saturated thickness of aquifer remains unchanged)

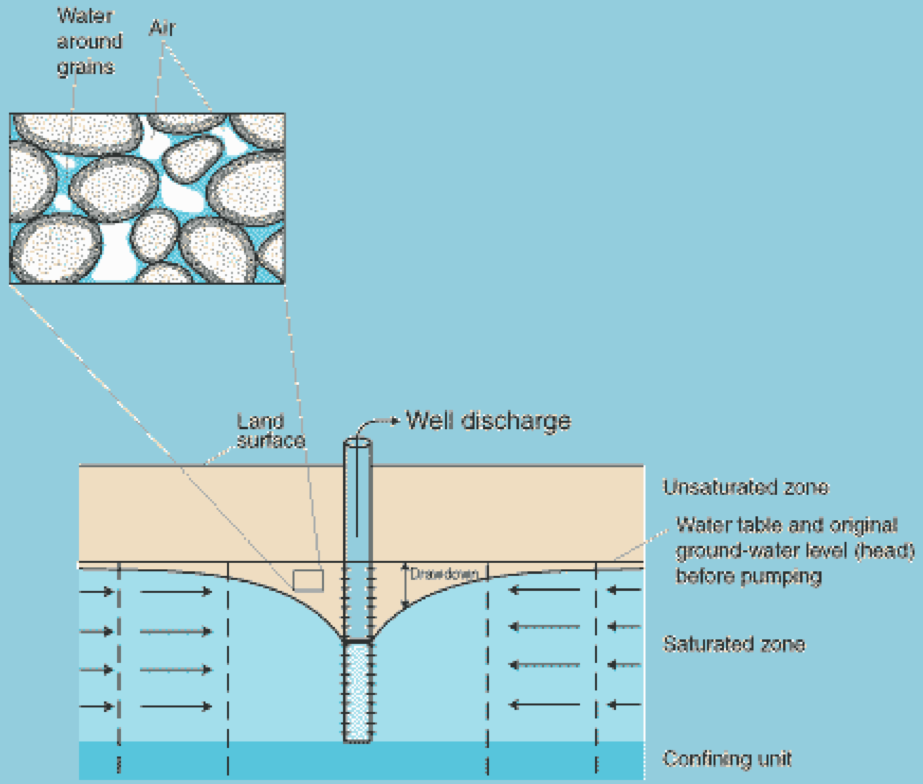


Fig A2:- Pumping a single well in an idealized unconfined aquifer. Dewatering occurs in cone of depression of unconfined aquifers during pumping by wells (saturated thickness of aquifer decreases).

Release of water from confined and unconfined aquifers (USGS Illustration)

- The large differences in drawdowns and related volumes of the cone of depression in the two types of aquifers relate directly to how the two types of aquifers respond to pumping.

- In unconfined aquifers (Figure A-2) dewatering of the formerly saturated space between grains or in cracks or solution holes takes place. This dewatering results in significant volumes of water being released from storage per unit volume of earth material in the cone of depression.

Release of water from confined and unconfined aquifers (USGS Illustration)

- On the other hand, in confined aquifers (Figure A-1) the entire thickness of the aquifer remains saturated during pumping. However, pumping causes a decrease in head and an accompanying decrease in water pressure in the aquifer within the cone of depression. This decrease in water pressure allows the water to expand slightly and causes a slight compression of the solid skeleton of earth material in the aquifer.

- The volume of water released from storage per unit volume of earth material in the cone of depression in a confined aquifer is small compared to the volume of water released by dewatering of the earth materials in an unconfined aquifer. The difference in how the two types of aquifers respond to pumping is reflected in the large numerical difference for values of the storage coefficient *S*

Water levels in unconfined & confined aquifers

Phreatic water levels in unconfined aquifers are affected by direct recharge by rainfall, seepage from canals & reservoirs, recharge from or discharge to streams, withdrawal of water from wells and sometimes changes in atmospheric pressure.

Piezometric water levels in confined aquifers are affected by surface water stages, surface loading, changes in atmospheric pressure and earthquakes.

Pressure Effects on Water Levels

Water levels in confined aquifers are affected by fluctuations in atmospheric pressure. As the atmospheric pressure increases, the water level falls and vice versa. The ratio of change of water level in a well to change in atmospheric pressure is known as Barometric efficiency (BE) and it is usually expressed as a percentage

BE= Δh/(Δpa)

Δh = *change in* piezometric *water level resulting from a change in atm.pressure in meters*

Δpa = *change in barometric pressure in meters of water*

Pressure Effects on Water Levels

Theoretically, a well with a full response to changes in barometric pressure would have a barometric efficiency of 100 percent and a well unaffected by changes in barometric pressure would have a barometric efficiency of 0 percent. Typically, barometric efficiency values range from 20 to 70 percent (Todd 1980).

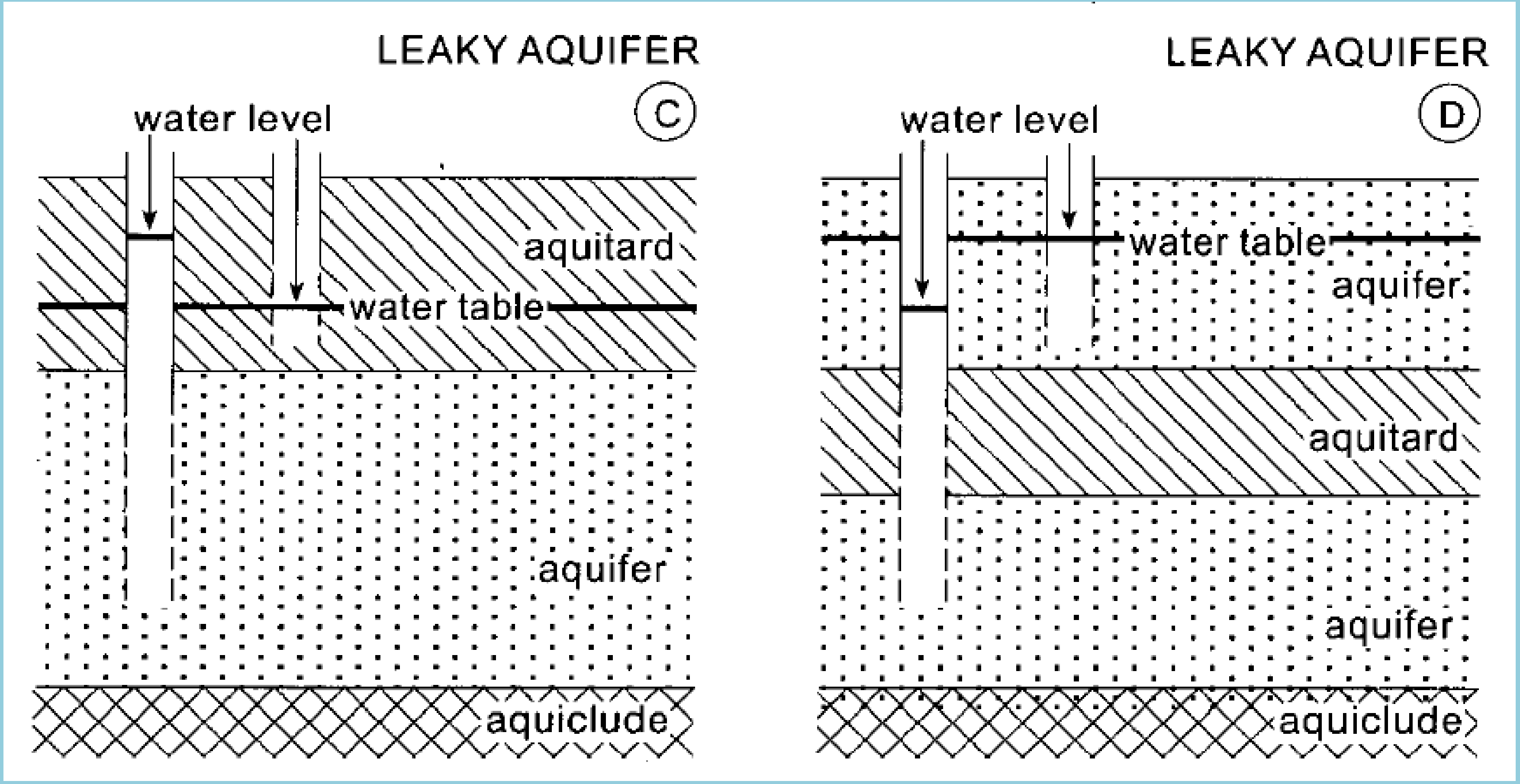
LEAKY or SEMI-CONFIEND AQUIFERS

- **A leaky aquifer, also known as a semi-confined aquifer, is a completely saturated aquifer that is bounded above by an aquitard and below by an aquitard/aquiclude.**
- **The aquitard are semi- permeable, it may slowly transmit appreciable water to or from adjacent aquifers.**
- **Confined aquifers that loose or gain waters from the surrounding formations are called *leaky confined aquifers.***

LEAKY or SEMI-CONFIEND AQUIFERS

- **If the overlying aquitard extends to the land surface, it may be partly saturated (Figure C), but if it is overlain by an unconfined aquifer that is bounded above by the water table (Figure D), it will be fully saturated.**

LEAKY or SEMI-CONFIEND AQUIFERS



LEAKY or SEMI-CONFIEND AQUIFERS

If there is hydrological equilibrium, the piezometric water level in a well tapping a leaky aquifer may coincide with the water table.

In areas with upward or downward flow, in other words, in discharge or recharge areas, the piezometirc level may rise above or fall below the water table.

In semi-confined aquifers the preferred direction of flow in the confining layers above and below is vertical.

The permeability of aquitard is less as compared to permeability of main aquifer.

SEMI-UNCONFIEND AQUIFERS

If however, the permeability of the main aquifer is not too great to ignore the horizontal flow component in the covering layer such an aquifer is intermediate between the traditional semi-confined aquifer and the unconfined aquifer and may be termed as ‘semi-unconfined aquifer’.

Semi-unconfined aquifers are aquifers which exhibit characters in between semi-confined and unconfined aquifers.

Here the permeability of the fine grained overlying layers is more than in a semi-confined aquifer and the horizontal flow component in it cannot be neglected.

PERCHED AQUIFERS

- **It is a special type of an unconfined aquifer.**
- **Sometimes, an impermeable bed of clay or silt may be present in some areas above the regional water table within the vadose zone or zone of aeration.**
- **This impermeable barrier intercepts downward movement of water and causes some of it to accumulate in the interstices of the rocks present above the stratum.**

PERCHED AQUIFERS

- **Thus, a zone of saturation of limited areal extent is locally formed with in the zone of aeration i.e. a small water-bearing zone sometimes exists between the main water table and the ground surface.**
- **This zone is called the perched ground water zone and the aquifer is called a *perched aquifer*. The upper surface of the ground water in this case is called a *perched water table*.**
- **The perched aquifer has limited thickness and areal extent.**

Classification of aquifers based on the Permeability of the Covering Layer

- **Based on the permeability of the covering layer Kruseman and De Ridder (1970) have given the following distinguishing features of the different type of aquifers.**

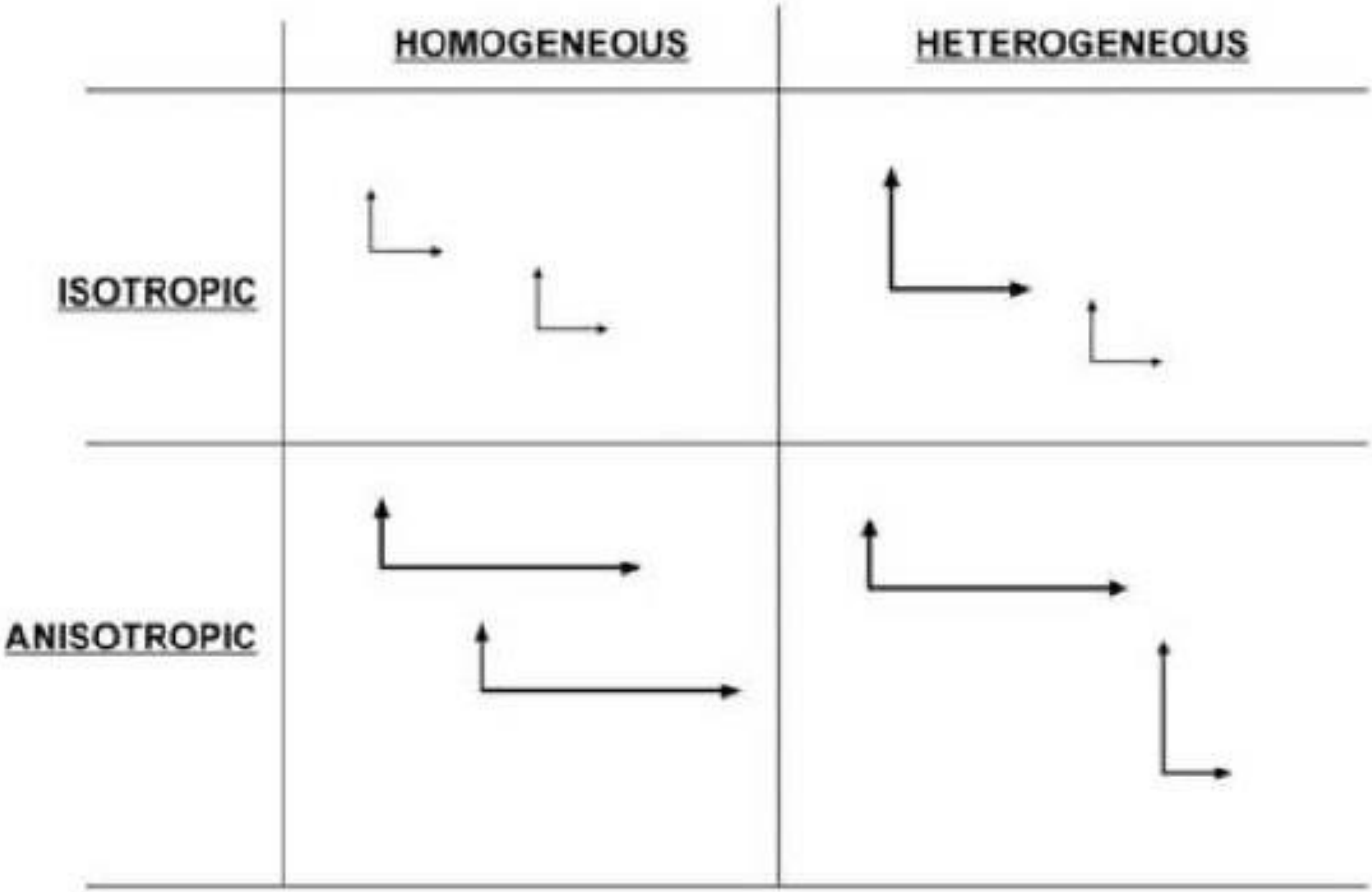
Covering Layer	Aquifer type
Impervious	Confined
Semi-pervious, so that the horizontal flow can be neglected	Semi-confined
Same as the main part of the aquifer	Unconfined

Classification Aquifers based on permeability

- **The hydraulic conductivity may vary spatially.**
- **If the K is essentially same throughout the geological formation, the aquifer is said to *homogeneous*.**
- **If it is different in different locations, then it is said to be *heterogeneous*.**

Classification Aquifers based on hydraulic conductivity (permeability)

- **The hydraulic conductivity can also vary with respect to directions.**
- **If the hydraulic conductivity is essentially the same in all directions, the aquifer is said to be *isotropic* .**
- **If it is different in different directions, the aquifer is said to be *anisotropic* .**


HOMOGENEOUS
HETEROGENEOUS
ISOTROPIC
ANISOTROPIC

Well Hydraulics

When you pump a well,

- **water level goes down**
- **but not the same way as in case of a tank**
- **you can see it in a dug well**
- **and measure it in the well**
- **or in a nearby observation well /Piezometer**

- **This difference is because of**
  - **Properties of the fluid (water) and**
  - **properties of the medium (aquifer)**

Properties of the fluid (water)

- **mass density of fresh water (*ρw)***
- **Weight density (*ρwg)***
- **compressibility (*β)*Water is often considered incompressible, but it does have a finite, low compressibility**
- **dynamic viscosity (*μ)***

Properties of Medium

(AQUIFER PROPERTIES)

AQUIFER PROPERTIES

- Aquifer performs two functions namely
- Storage Function
- Conduit/Transmitting Function.

Storage Properties of the Aquifer

Storage Properties

- Porosity
- Specific Storage
- Storage Coefficient
- Specific Yield

POROSITY

- It as the ratio of the volume of voids to the total volume and can be expressed as a percentage or as decimal fraction.
- Porosity (n) = $\frac{\text{Void Volume}}{\text{Total Volume}}$ n= $\frac{Vv}{VT}$ X 100

POROSITY

- ***Primary porosity*** **is the inherent character of a rock which is developed during the formation of the rock itself.**
- **In sedimentary rocks & alluvial formations – it is the inter-granular space.**
- **In volcanic igneous rocks, the primary porosity is due to the presence of gas cavities (vesicles) and also lava tubes and lava tunnels.**

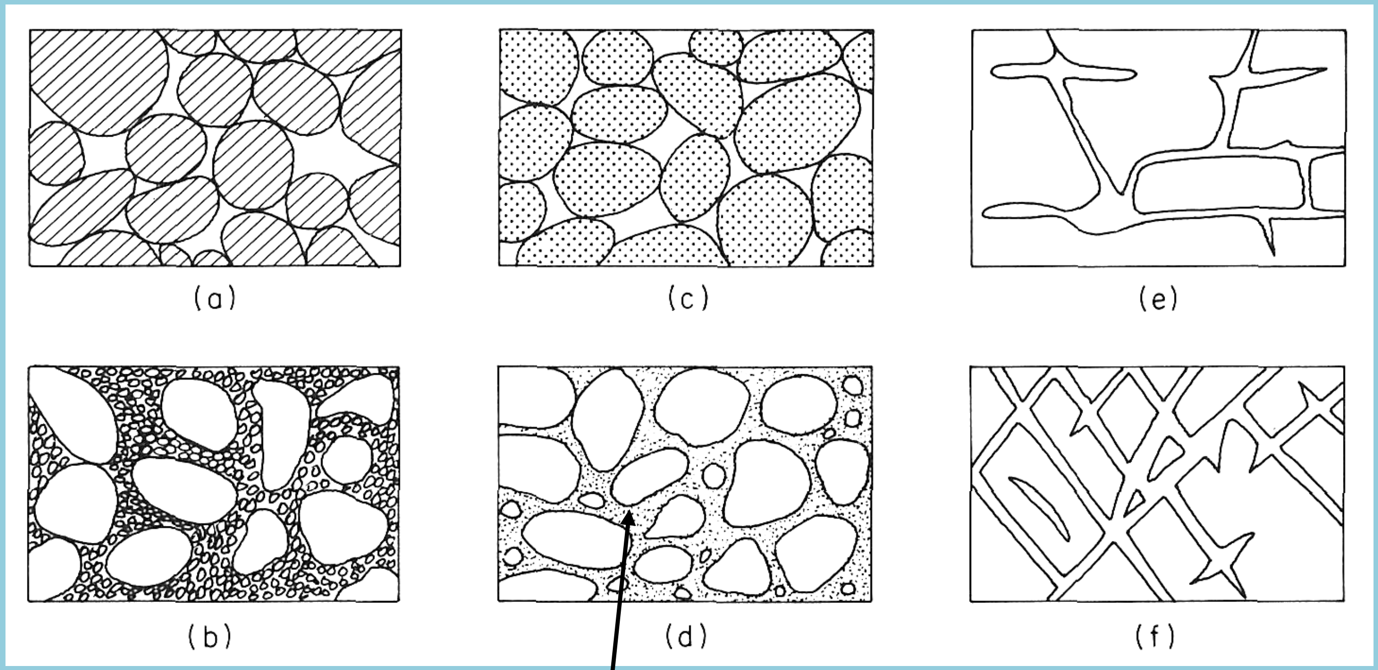
POROSITY

Secondary porosity **is the induced character and is developed subsequent to the formation of rocks.**

e.g. weathering, joints and fractures, dissolution of minerals like in carbonate rocks.

In sedimentary rocks, primary porosity plays a significant role.

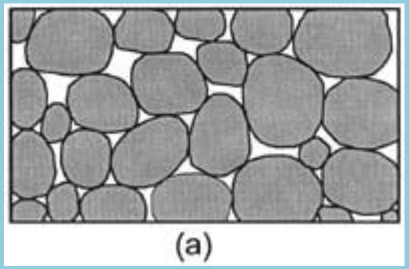
Types of Porous Media



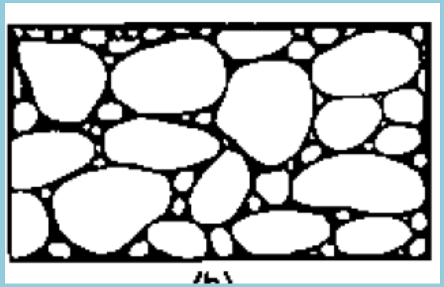
$$n = V_V / V_T = \text{Vol Voids / Total Vol}$$

Freeze and Cherry, 1979.

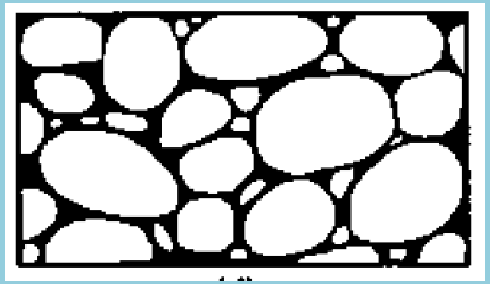
Rock texture and porosity



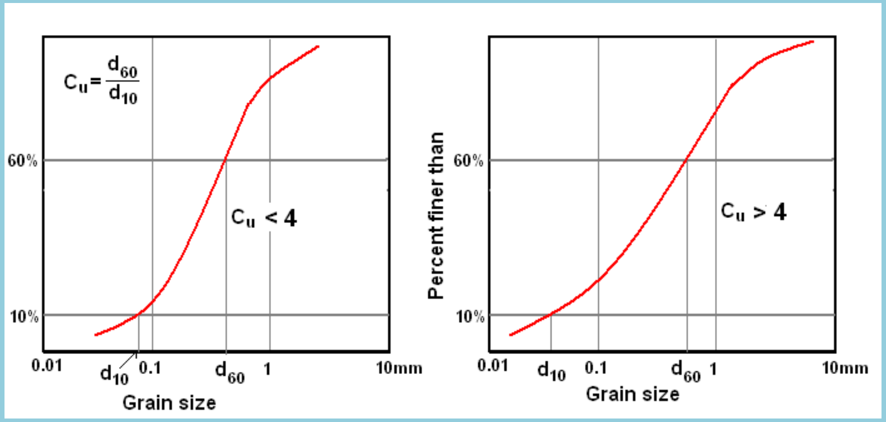
Well sorted sedimentary deposit with high porosity



Poorly sorted sedimentary deposit with low porosity



Porosity reduced by cementation

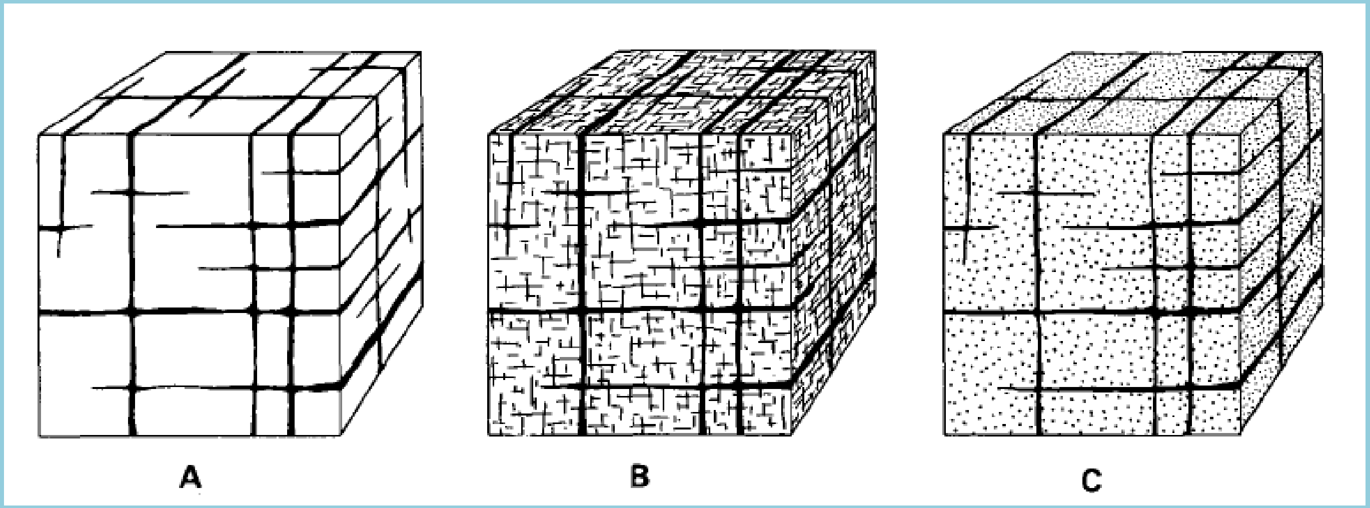


Relation between Packing & Porosity

(a) Square packing ($n = 47.6\%$) (Loosest state of packing)

(b) Rhombic packing ($n = 26\%$)

Hard-rock Porosity



Single porosity Microfissures Double porosity

Range of Values of Porosity (after Freeze & Cherry, 1979)

Unconsolidated deposits	
Gravel	25 - 40
Sand	25 - 50
Silt	35 - 50
Clay	40 - 70
Rocks	
Fractured basalt	5 - 50
Karst limestone	5 - 50
Sandstone	5 - 30
Limestone, dolomite	0 - 20
Shale	0 – 10
Fractured crystalline rock	0 - 10
Dense crystalline rock	0 – 5

Effective porosity

- \- the amount of interconnected pore spaces available for fluid flow.
- it is the ratio of volume of interconnected voids to total volume of rocks.
- $$ne = \frac{\text{Volume of interconnected voids}}{\text{Total volume}}$$

Porosity and effective porosity

- These two porosities are not identical. For example, many crystalline rocks have a high total porosity, most of which may be unconnected.
- Effective porosity is less than the overall total porosity. (Clays porosity is 60% but effective porosity is 3%)

Specific Storage (Ss)

- **The specific storage of a saturated confined aquifer is the volume of water that a unit volume of aquifer releases from storage under a unit decline in hydraulic head. The specific storage is defined as:**

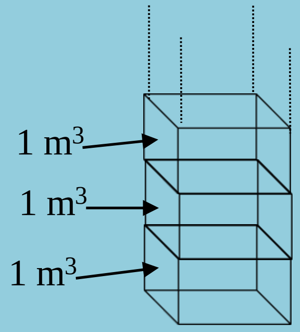
$$S_s = \rho g(\alpha + n\beta)$$

- **where α is compressibility of the porous medium value of which ranges from 10^{-6} to 10^{-8} for clays, 10^{-7} to 10^{-9} for sands and from 10^{-8} to 10^{-10} for fractured rocks, its unit is m^2/N or Pa^{-1}.**
- **β is the compressibility of water and is usually taken as 4.4×10^{-10} m^2/N or (Pa^{-1}).**

Specific Storage (Ss)

- **S_s = volume of water released per 1 m^3 of aquifer volume per 1m change in head.**
  example: S_s = 0.01 m^{-1} (0.01 m^3 released per 1m^3 box)

- **i.e., if head drops 1m while 0.01 m^3 is released**



$$S_s = \frac{\left(\dfrac{0.01m^3}{1m^3}\right)}{1m} = 0.01m^{-1}$$

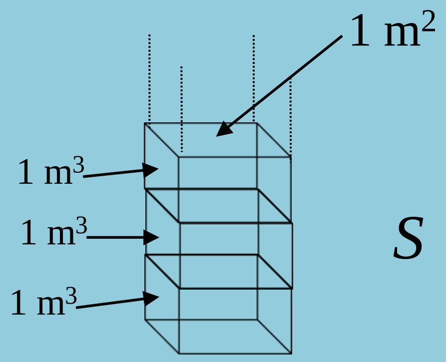
CO-EFFICIENT OF STORAGE

- The capacity of an aquifer to store water is expressed as a coefficient designated as S.
- The head in the aquifer changes when water is either stored or released indicating a change in the storage volume within the aquifer.

Storativity or Storage Coefficient (S)

volume of water released or taken into storage per unit of aquifer storage area per unit change in head.

- Aquifer volume per 1 m^2 of aquifer area = aquifer thickness = b
  - example: = 3m x 1m^2 / 1m^2 = 3m = b



$$S = \frac{\frac{V_{released}}{\text{area}}}{\text{change in head}} = \frac{\frac{3 \times 0.01 m^3}{1 m^2}}{1 m} = 0.03 = S_s \cdot b$$

Storativity or Storage Coefficient (S)

- **Storativity is defined as the volume of water which a vertical column of an aquifer releases from storage or takes in to storage per unit surface area of the aquifer per unit change in component of the head normal to that surface.**

$$S = \frac{\text{volume of water}}{(\text{unit area})(\text{unit head change})} = \frac{(m^3)}{(m^2)(m)} = \frac{m^3}{m^3}$$

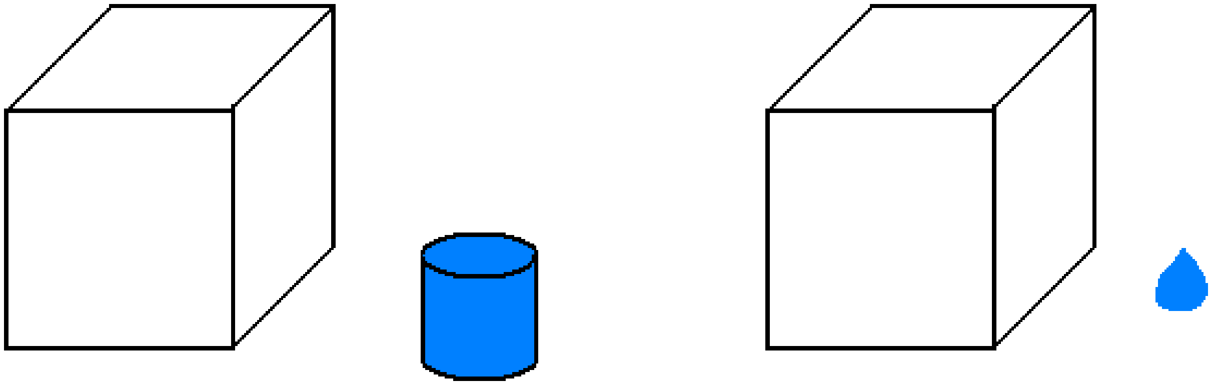
- $S = S_s D$

Storativity or Storage Coefficient (S)

- In confined aquifer, storativity is a result of compressibility of the aquifer and expansion of the contained water a result of reduced pressure due to pumping. The value of S ranges from 0.00001 (10^{-5}) to 0.001 (10^{-3}) for confined aquifers.
- Storativity is non-dimensional.

Un-confined Aquifer

Confined Aquifer

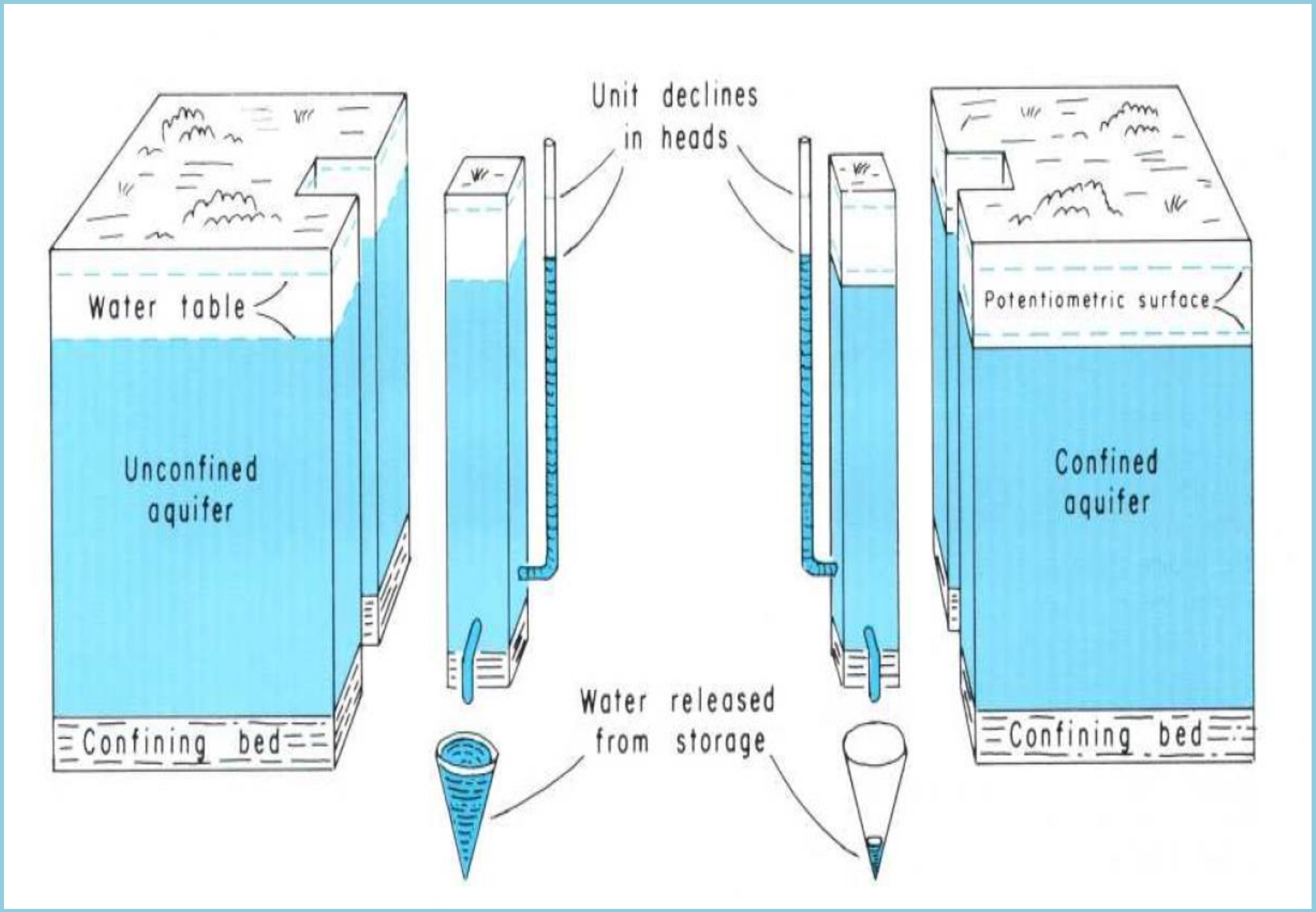


Specific yield (Sy)

0.02 - 0.25

Storage Coefficient (S)

$5\times10^{-3} — 5\times10^{-5}$


Unit declines
in heads
Water table
Unconfined
aquifer
Confining bed
Water released
from storage
Potentiometric surface
Confined
aquifer
Confining bed

SPECIFIC YIELD (Sy)

- **In case of an unconfined aquifer the concept of storativity is analogous to that of specific yield.**

- **In an unconfined aquifer, the water released from storage is mainly due to gravity drainage and not due to the compressibility of aquifer material or of water.**

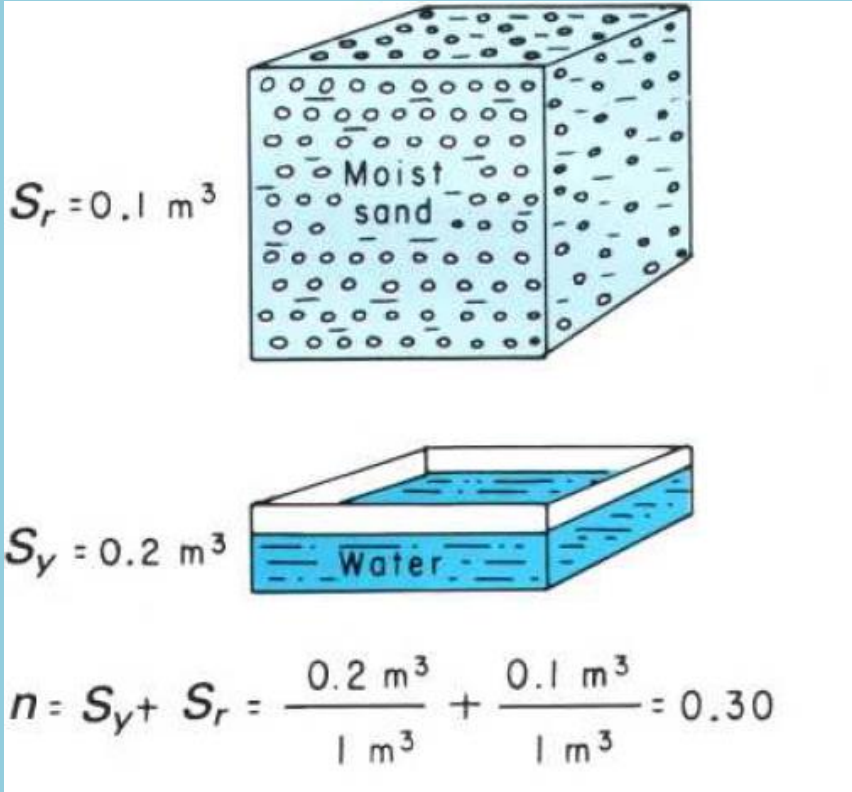
SPECIFIC YIELD (Sy)

- **When water is drained from a saturated material by gravity force, only part of the total volume stored in its pores is released. The quantity of water that a unit volume of material will give up when drained by gravity is the specific yield.**

Specific Retention

- **The part of water that is not removed by gravity drainage is held against the force of gravity by molecular attraction and capillary. The quantity of water that a unit volume of aquifer retains when subjected to gravity drainage is called its *specific retention.***

Specific yield and specific retention

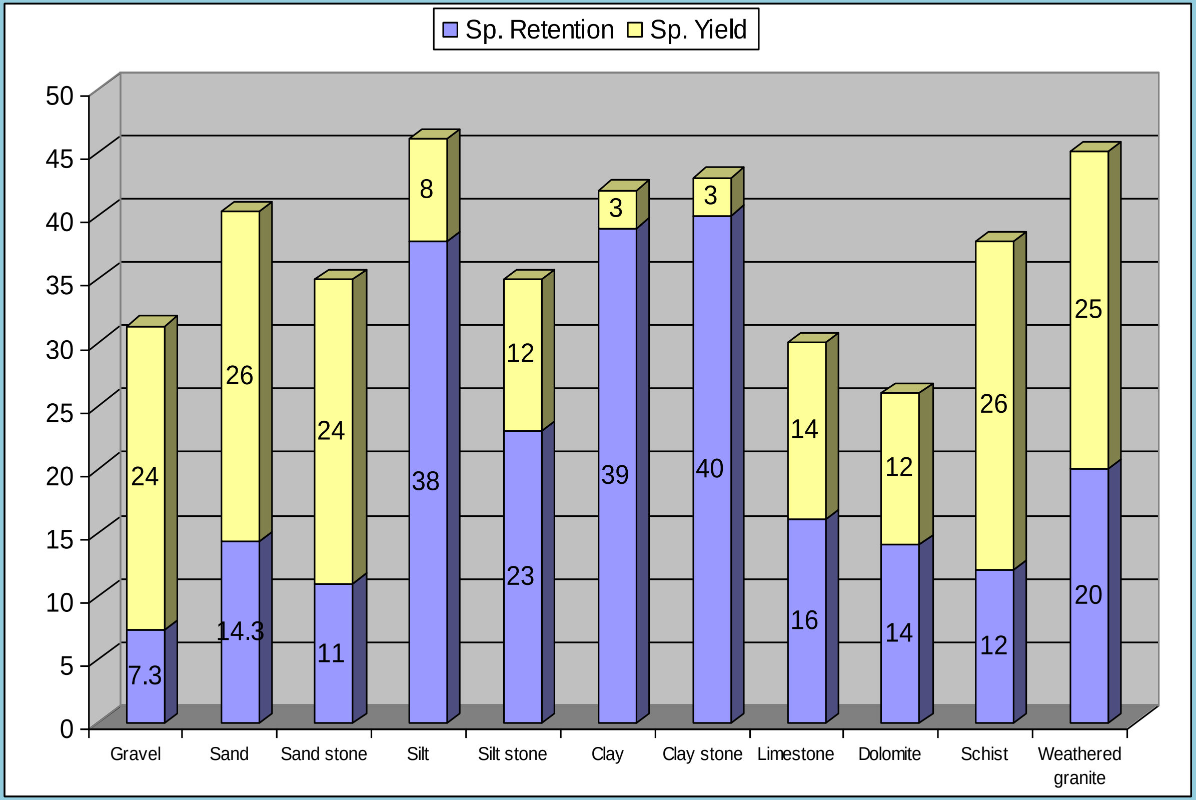


- **For e.g. if 0.20 m^3 of water is drained from one cubic meter of saturated sand, the specific yield of sand is 0.20 or 20%.**
- **The specific yield in unconfined aquifers ranges from 0.05 to 0.30.**

Specific yield and specific retention

- **The specific yield and specific retention depend upon the shape and size of particle, distribution of pores (voids), and compaction of the formation.**
- **The specific retention increases with decreasing grain size.**
- **It should be noted that it is not necessary that soil with high porosity will have high specific yield because that soil may have low permeability and the water may not easily drain out.**
- **For example, clay has a high porosity but low specific yield and its permeability is low.**

Porosity, Specific Yield, Specific Retention



Specific yield values recommended by GEC-1997

S.No	Formation	Recommended Value (%)	Minimum (%)	Maximum (%)
(a)	Alluvial areas			
	Sandy alluvium	**16.0**	12.0	20.0
	Silty alluvium	**10.0**	8.0	12.0
	Clayey alluvium	**6.0**	4.0	8.0
(b)	Hard rock areas			
	Weathered granite, gneiss and schist with low clay content	**3.0**	2.0	4.0
	Weathered granite, gneiss and schist with significant clay content	**1.5**	1.0	2.0
	Weathered or vesicular, jointed basalt	**2.0**	1.0	3.0
	Laterite	**2.5**	2.0	3.0
	Sandstone	**3.0**	1.0	5.0
	Quartzite	**1.5**	1.0	2.0
	Limestone	**2.0**	1.0	3.0
	Karstified limestone	**8.0**	5.0	15.0
	Phyllites, Shales	**1.5**	1.0	2.0
	Massive poorly fractured rock	**0.3**	0.2	0.5

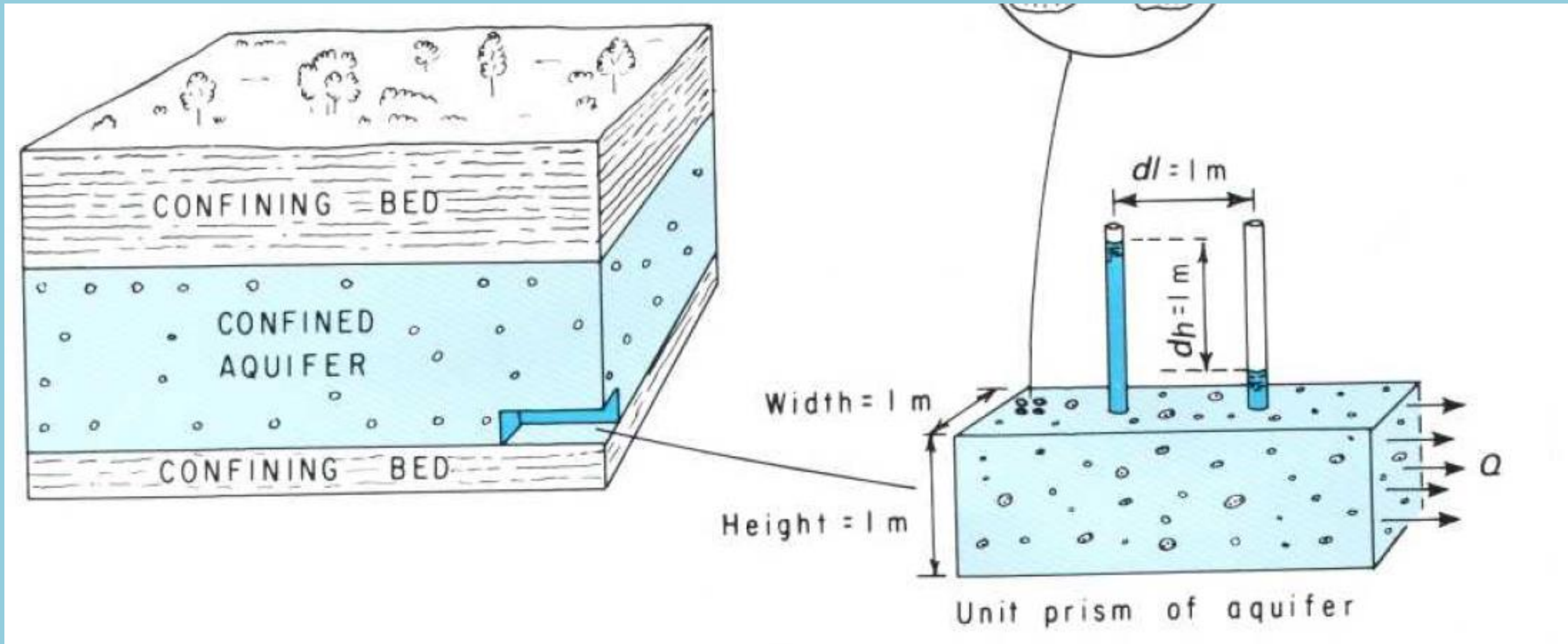
Transmitting Properties of the Aquifer

Transmitting Properties of the Aquifer

- **Hydraulic Conductivity or coefficient of permeability or simply Permeability (K)**
- **Transmissivity (T)**

Hydraulic Property: PERMEABILITY

- **Permeability or coefficient of permeability (K) is the ability of a porous medium to transmit a fluid or water.**
- **It is measured by the rate at which it will transmit water through a given cross section under a given difference of pressure per unit of distance.**



discharge (m^3/s) per unit area (m^2) of rock/soil mass under unit hydraulic gradient.

Hydraulic Property: PERMEABILITY

- **Because the discharge per unit area equals to the velocity, the coefficient of permeability has the dimension of the velocity [L/T].**

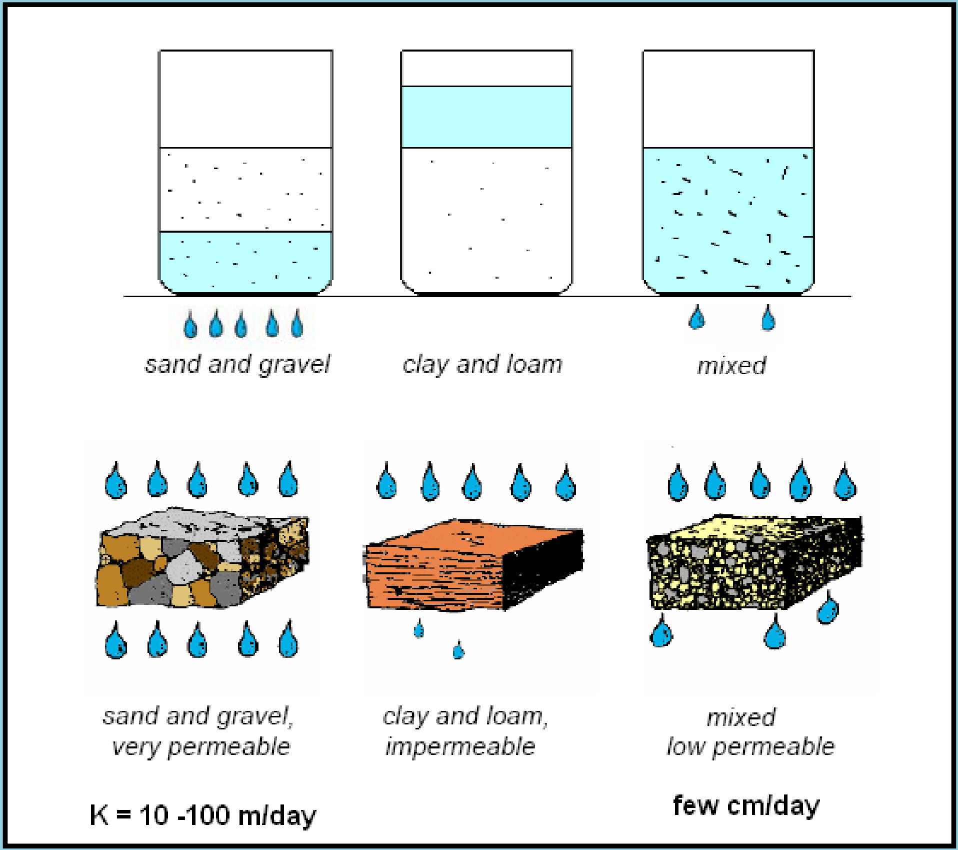
$$Q = Av. \text{ or } v = \frac{Q}{A}$$

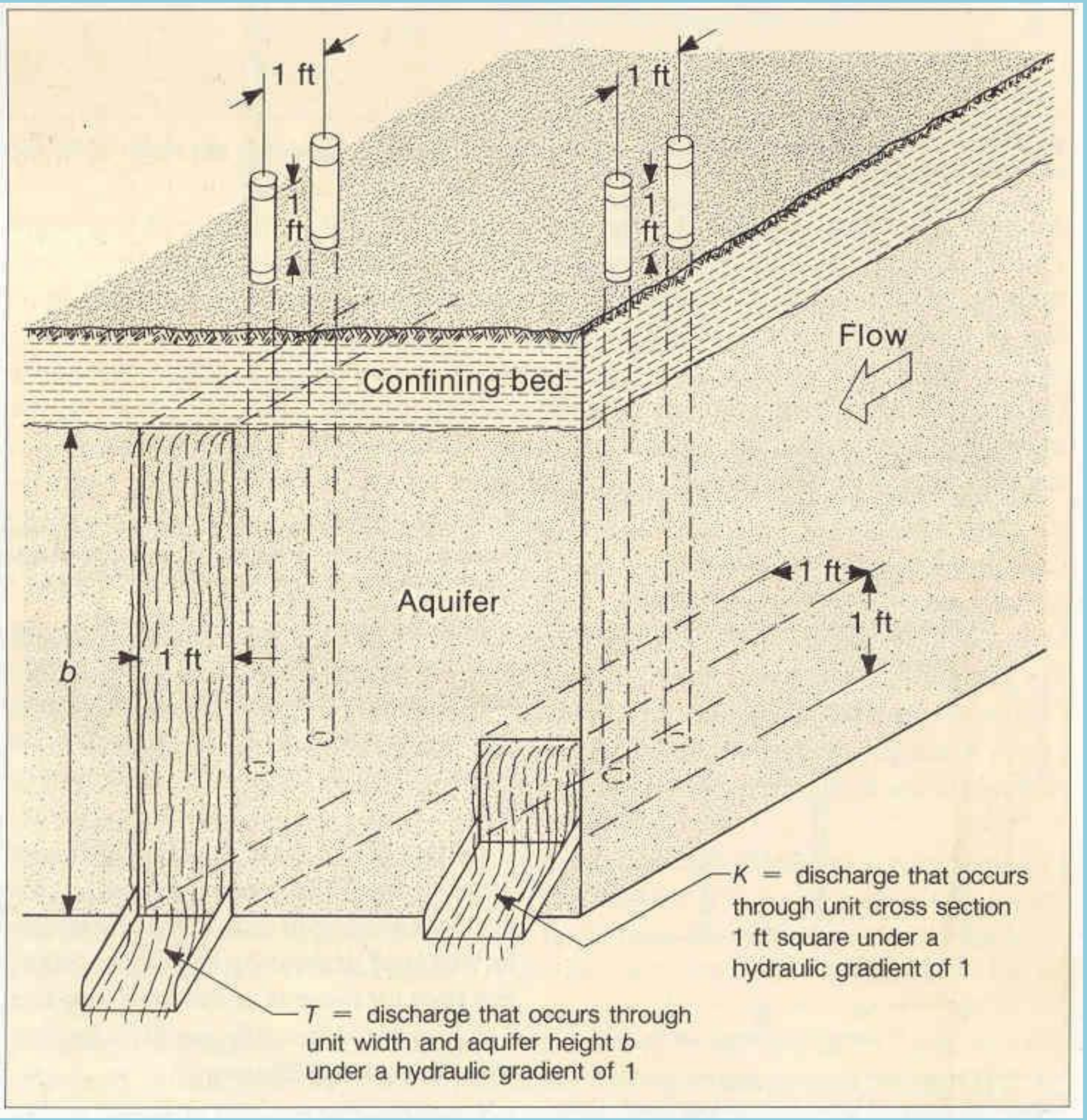
- **It is usually expressed as cm/s, m/s, m/day, etc.**

- **The coefficient of permeability is also known as Hydraulic Conductivity.**

Permeability

The ability of a formation to transmit water




1 ft
1 ft
1 ft
1 ft
Confining bed
Flow
Aquifer
1 ft
1 ft
1 ft
b
K = discharge that occurs through unit cross section 1 ft square under a hydraulic gradient of 1
T = discharge that occurs through unit width and aquifer height b under a hydraulic gradient of 1

TRANSMISSIVITY

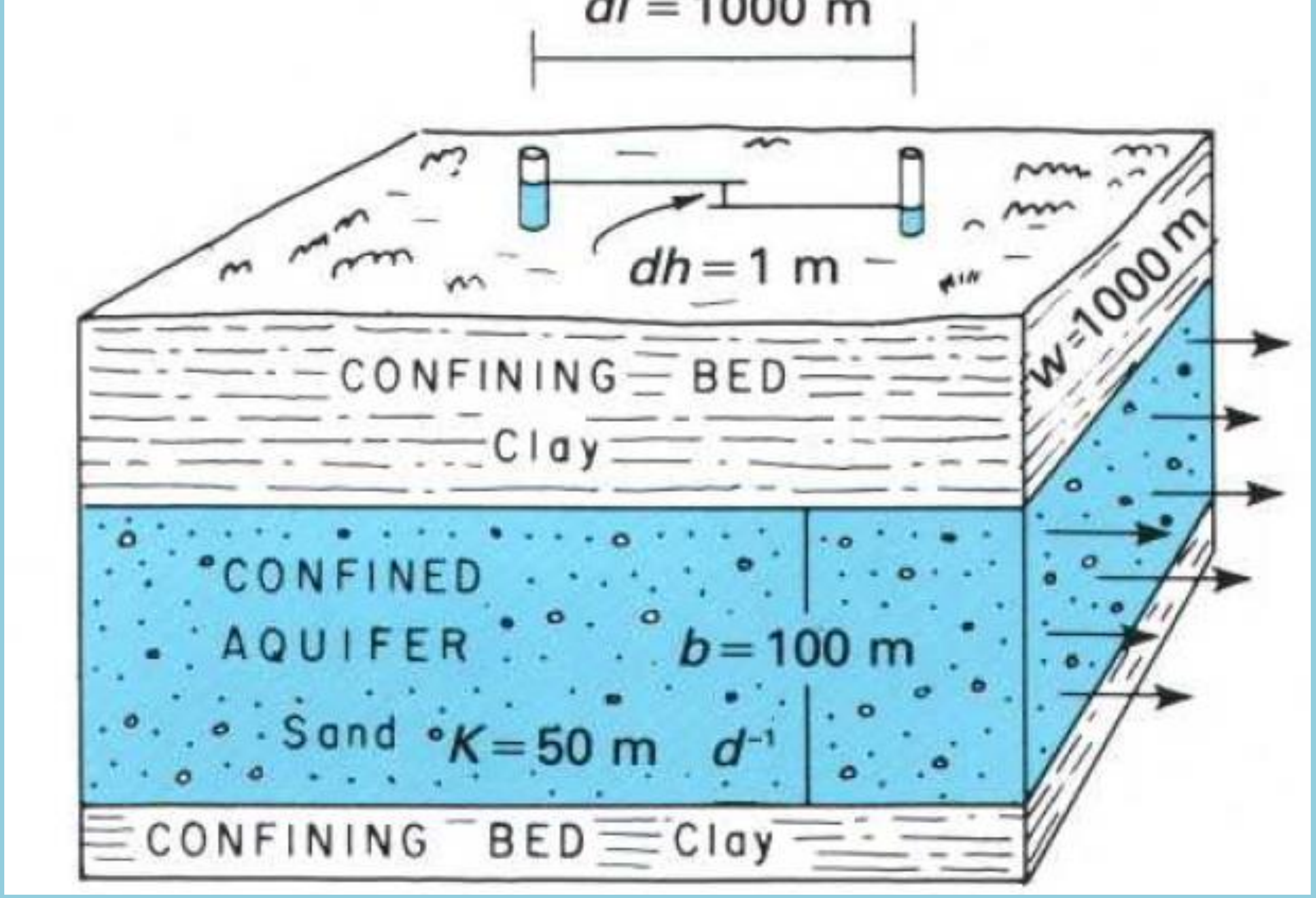
- The capacity of an aquifer to transmit water depends on the thickness and hydraulic conductivity of the aquifer.
- It is a product of average hydraulic conductivity and saturated thickness of the aquifer.
- $T = Kb$
- (T= transmissivity in m^2/day, K= hydraulic conductivity in m/day, b = thickness of the aquifer in m.)

Hydraulic Conductivity for Unconsolidated and Consolidated Rocks (m/day)

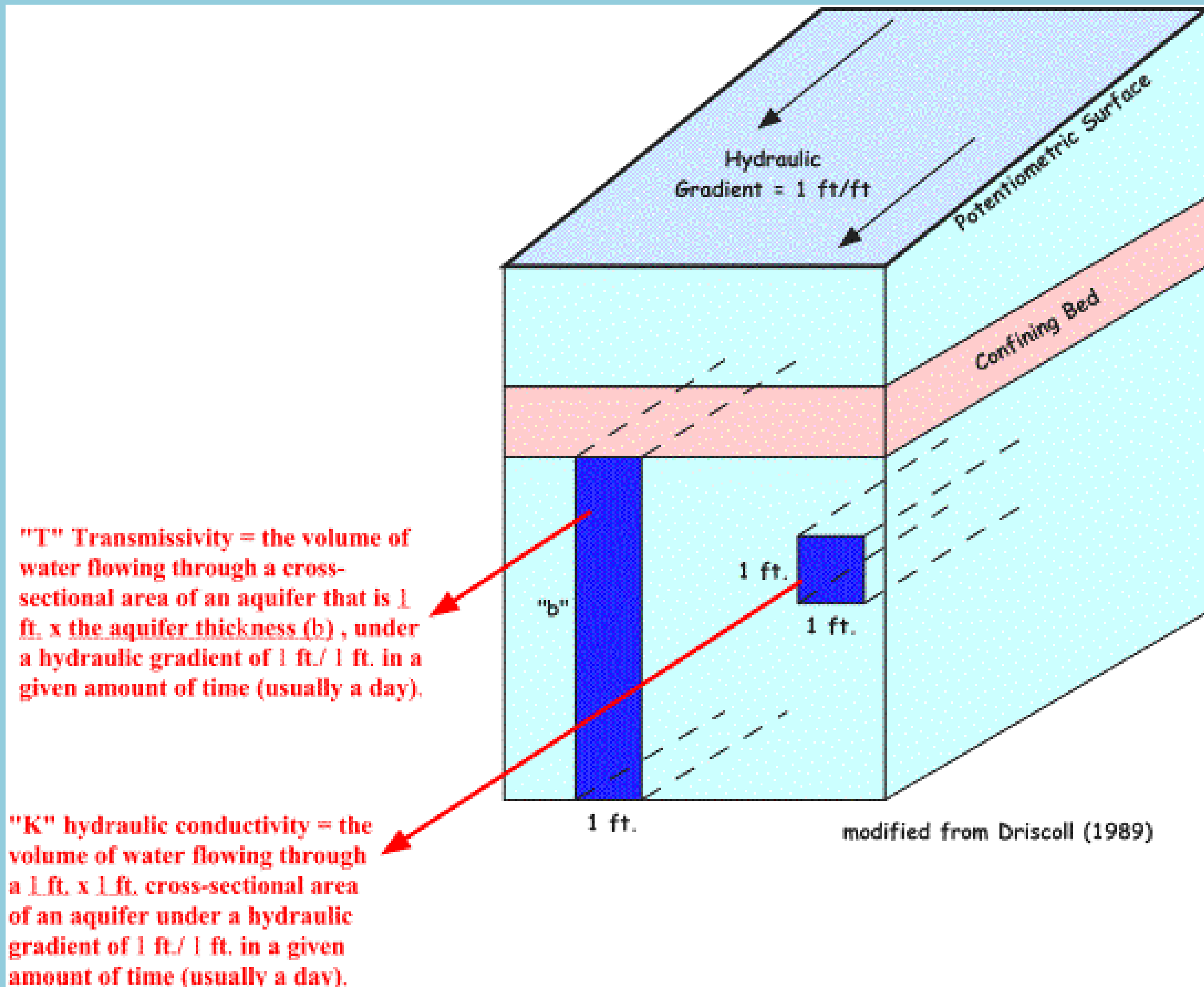
Unconsolidated deposits	
Clay	$10^{-8} – 10^{-2}$
Fine sand	1 - 5
Medium sand	5 - 20
Coarse sand	$20 - 10^{2}$
Gravel	$10^{2} - 10^{3}$
Sand and gravel mixes	$5 - 10^{2}$
Clay, sand, gravel mixes (e.g. till)	$10^{-3} – 10^{-1}$
Hard Rocks	
Chalk (very variable according to fissures if not soft)	30.0
Sandstone	3.1
Limestone	0.94
Dolomite	0.001
Granite, weathered	1.4
Schist	0.2

TRANSMISSIVITY

Transmissivity is defined as the rate of flow of water in meters square per day, through a vertical strip of the aquifer of one meter wide (unit width) and extending through the entire saturated thickness of the aquifer under a hydraulic gradient of (100% unit hydraulic gradient) at a temperature of 15.6°C.



dl = 1000 m
dh = 1 m
w = 1000 m
CONFINING BED
Clay
CONFINED
AQUIFER
b = 100 m
Sand
K = 50 m d⁻¹
CONFINING BED Clay



Transmitting capacity through unit cube of aquifer – ‘K’
Transmitting capacity through unit prism of aquifer – ‘T’

Hydraulic Diffusivity (a)

- **The ratio of transmissivity to the coefficient of storage of an aquifer is defined as its hydraulic diffussivity.**

  a = (T/S)

- **This parameter determines the time that is needed for a given head change to occur in an aquifer in response to greater change in head at another point.**
- **It is the rate of propagation of change in head in an aquifer and is given by T/S.**
- **hydraulic diffusivity unit = [m2/d]**

Properties of the Semi-confining layer

Leakage Coefficient or Leakance (Lc)

- This is the property of the semi-confining (aquitard) layer and is a measure for the vertical conductivity of a semi-confining layer.
- It is the ratio of vertical permeability of semi-confining layer to its thickness. i.e

$$Lc = \frac{K'}{D'}$$

- **(Dimensions T−1)**

- It is used to characterize the amount of leakage through an aquitard.

Hydraulic Resistance (c)

It defines the resistance of the semi-confining layer against vertical flow. It is the reciprocal of leakage coefficient.

$$c = \frac{D'}{K'}$$

- It characterises the resistance of the semi-pervious layer (aquitard) to upward or downward leakage, it has dimension of time and its unit is generally days. Values of c vary from some hundreds of days to several ten thousand days.
- For aquicludes, c is infinite. If the hydraulic resistance c = ∞, the aquifer is confined.
- As this value becomes larger, the leakage through this layer diminishes.

Leakage factor (L)

The leakage factor is a measure for the spatial distribution of the leakage through an aquitard in to the leaky aquifer. It is defined as:

$$L = \sqrt{KDc}$$

where c is Hydraulic resistance of confining layer, K is Hydraulic conductivity of Leaky aquifer and D is its thickness.

High values of L indicates a great resistance of the semi-pervious strata to flow.

Leakage factor has dimension of length and is generally expressed in m.

Boulton's Delay Index (1/α)

- **This is a measure of the delayed drainage of an *unconfined aquifer.***
- **It has the dimensions of time. The value of 1/α may vary from about 50 min in coarse sand to 4000 min in silt and clay.**

Drainage Factor (B)

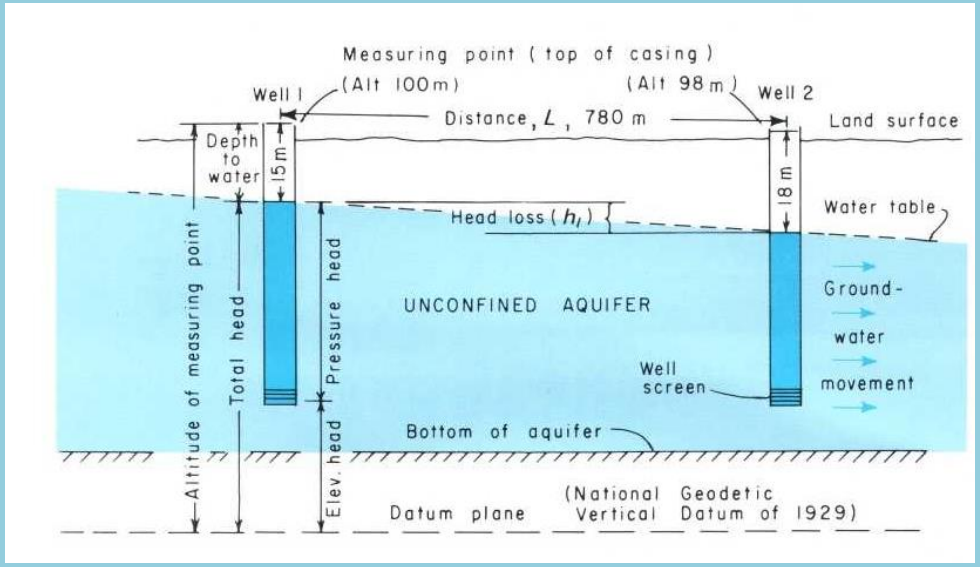
- **Drainage factor is a property of unconfined aquifer.**
- **($B = \sqrt{T/\alpha S_y}$) where** $1/\alpha$ = Boulton delayed index
- **It has the dimensions of length and is usually expressed in metres. Large values of B indicate fast drainage. If $B = \infty$, the yield is instantaneous with the lowering of the water-table, i.e. the aquifer is unconfined without delayed yield.**

What makes ground water move?

Differences in hydraulic head (the height to which water will rise in an open pipe)

HEAD AND GRADIENTS

❖ The equation for total head is $ht = Z + hp$
where z is elevation head and distance from the datum where pressure head hp is determined



❖ The hydraulic gradient is the change in head per unit of distance in a given direction . $= hl/L$ (Head loss/horizontal distance) 5/780

Ground Water Hydraulics

Driving Forces of Ground Water Flow Hydraulic head

In hydrogeologic practice the driving force for groundwater flow is generally expressed in terms of a parameter called *hydraulic head or simply head.*

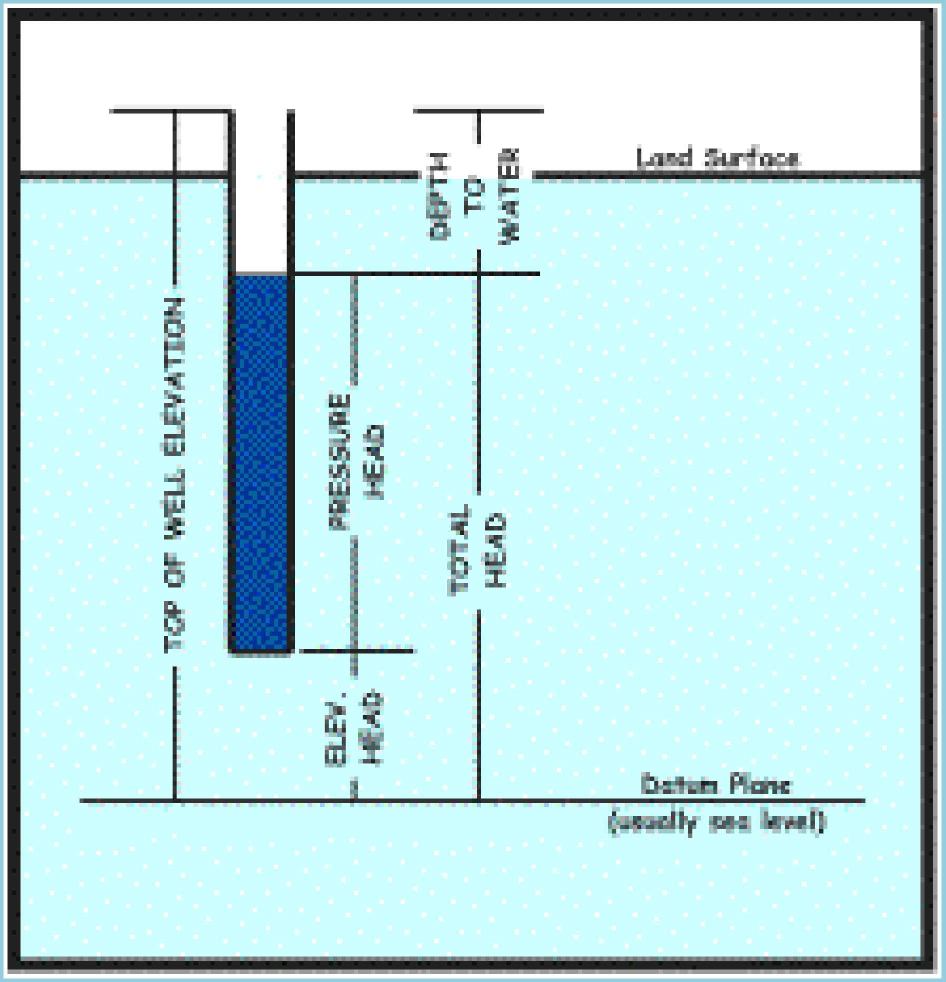
Ground Water Hydraulics

Driving Forces of Ground Water Flow (Hydraulic Head)

- **Hydraulic head or simply ‘head’ is a specific measurement of water level relative to a datum plane that is common to all the wells.**
- **In other words, it is the height that water will rise in a well relative to an arbitrary datum.**

Ground Water Hydraulics

Driving Forces of Ground Water Flow (Hydraulic Head)



Ground Water Hydraulics

Driving Forces of Ground Water Flow (Hydraulic Head)

- The hydraulic head consists of the following three components:-
- ***Elevation head (z)*** **is the distance of the bottom of a well above a datum –often sea level. (Potential energy).**
- ***Pressure head (h)*** **height of the water column in a well that exerts the pressure. For a fluid at rest, the Pressure at a point = weight of the overlying water per unit cross-sectional area:** $P=\rho g h p$**. (Pressure energy).**
- ***Velocity head*** $v^2/2g$**. (Kinetic energy).**

Ground Water Hydraulics

Driving Forces of Ground Water Flow (Hydraulic Head)

From *Bernoulli's equation*, the total energy at a given point in a fluid (H) is the sum total of potential energy (elevation energy) by virtue of its position from the height of the fluid relative to an arbitrary datum (elevation head/energy), plus pressure energy by virtue of its from pressure (analogous to the energy of the compressed string) in the fluid (pressure head/energy), plus kinetic energy associated with the movement of the fluid (velocity head). The sum total of these energy is a constant.

Ground Water Hydraulics

Driving Forces of Ground Water Flow (Hydraulic Head)

$$H = z + \frac{p}{\rho g} + \frac{v^2}{2g}$$

From the above it is seen that *the sum total of the above energy is a constant. This is known as Bernoulli's Principle.*

Driving Forces of Ground Water Flow (Hydraulic Head)

- **Because ground water moves relatively slowly, velocity head can be ignored.**

$$H = z + \frac{p}{\rho g}$$

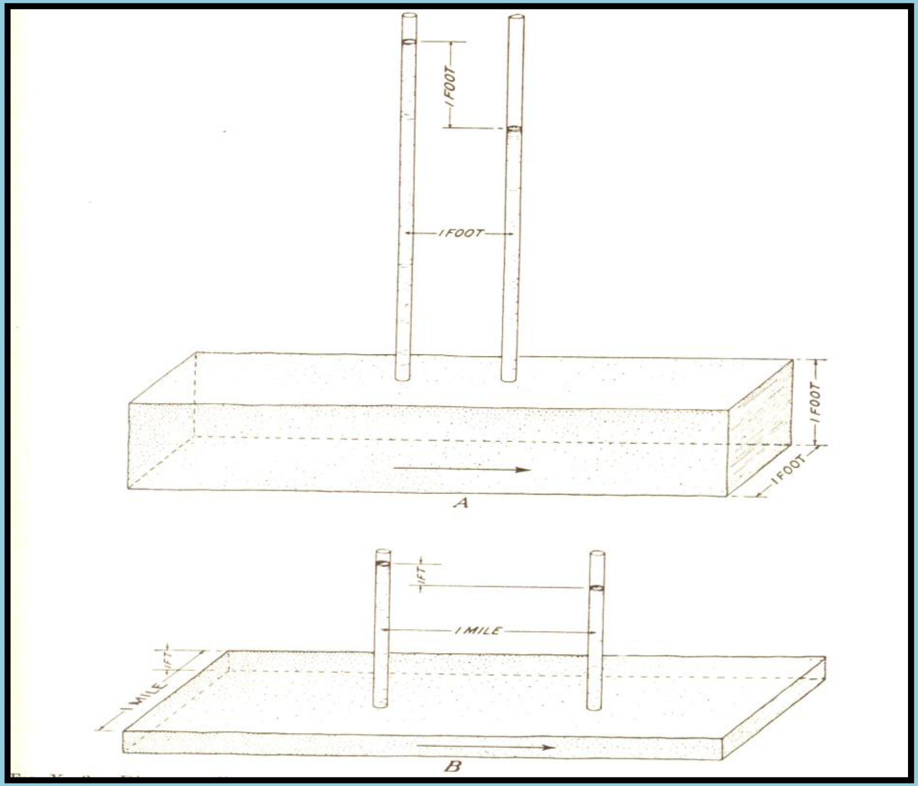
Hydraulic gradient (i)

- **Hydraulic gradient is the change in head over a distance.**
- **Hydraulic gradient, can be estimated from 3 or 4 wells.**

$$\textbf{Hydraulic Gradient (i)} = \frac{h_1 = \text{Up gradient Head [L]} - h_2 = \text{Down gradient Head [L]}}{\text{Distance Between Wells [L]}}$$

$$i = h_1 - h_2 / l \text{ or } i = dh/dl.$$

The *hydraulic gradient* is a vector gradient and it is also called the *Darcy slope.*



A hydraulic gradient of 1 means that the head falls 1 meter for every 1 meter of flow travel (one unit of length drop in unit of length of flow distance) as indicated in Fig

Ground Water Flow

- LAMINAR AND TURBULENT FLOW

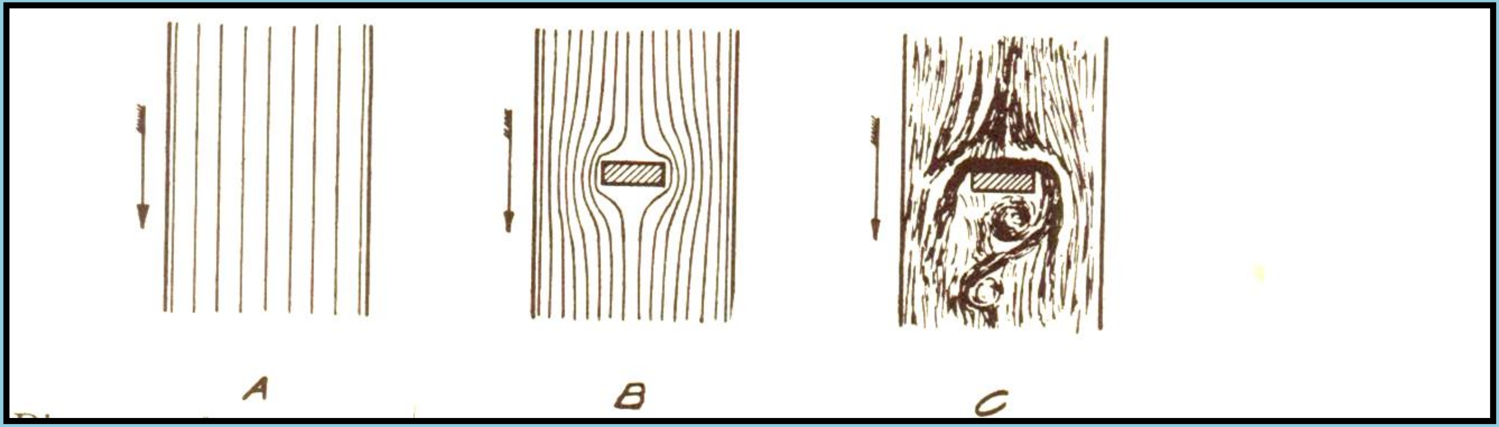
Two basic types of flow occur in ground water with one more prevalent than the other.

The water in the interstices of the permeable rocks in the zone of saturation is, as a rule, move very slowly and steadily. This slow and steady kind of movement is called the *laminar flow*. It is also known as *streamline (or) viscous flow*.

Ground Water Flow

- ***Turbulent flow or non-viscous flow*** **in which the movement of particles is tenuous and irregular, crossing and crisscrossing at random.**
- **In general, laminar flow occurs at relatively low velocities and turbulent flow at higher velocities. The flow in river and creeks is generally turbulent, whereas the flow of ground water is laminar.**

Ground Water Flow



LAMINAR AND TURBULENT FLOW

STEADY-STATE FLOW

- Flow is said to be under *steady* or *equilibrium* state when the magnitude and direction specific discharge remain constant with time.

It implies that the position of the piezometric surface and the hydraulic gradient remain unchanged. $\frac{dh}{dt} = 0$

- There is no addition to or withdrawal from the storage of the aquifer, and equilibrium conditions have been reached between recharge and discharge.

NON-STEADY STATE FLOW

- **Flow is said to be under *non-steady*, also called *unsteady* or *non-equilibrium* or *transient* state when the magnitudes or direction of specific discharge changes with time. Changes in storage of the aquifer are involved in non-steady flow. Non-steady state flow is described with respect to boundary and initial conditions.**

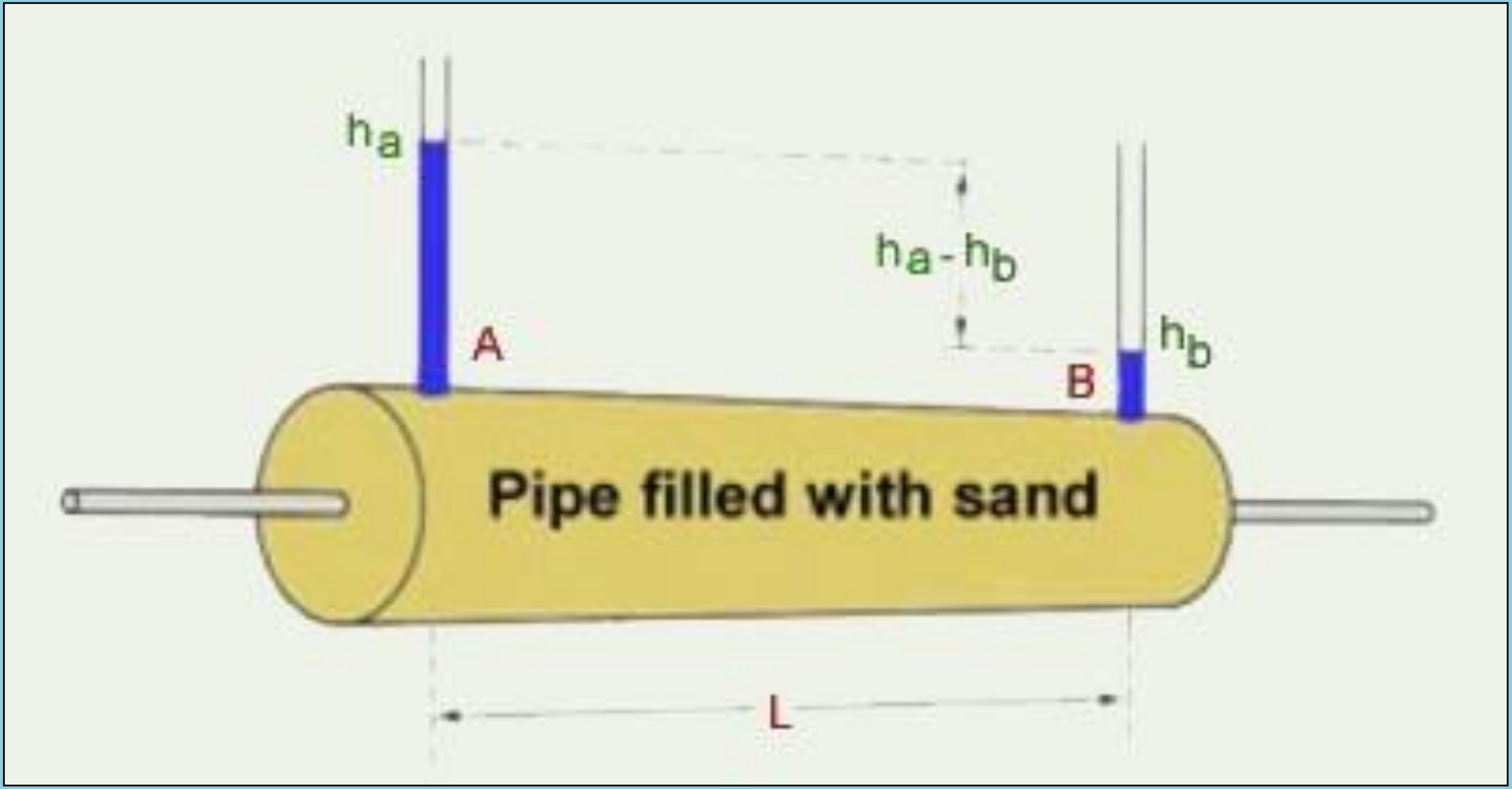
Darcy’s Law

- Darcy’s law provides an accurate description of the flow of ground water in almost all hydrogeologic environments.
- The law was formulated by Henry Darcy based on the results of experiments on the flow of water through beds of sand.

Darcy’s Law

- Published in 1856, his conclusions have served as the basis for all modern analysis of ground water flow

Darcy's Law



Darcy's Law

Conclusions:

- **Q is proportional to difference in height of the water ($h2 - h1$), known as 'head loss'**
- **Q is inversely proportional to the length of the flow path (L)**
- **Q is obviously proportional to the cross-sectional area of the pipe (A)**

Darcy's Law

Conclusions:

- ***Q*** **is proportional to difference in height of the water (*h*2 – *h*1), known as 'head loss'**
- ***Q*** **is inversely proportional to the length of the flow path (L)**
- ***Q*** **is obviously proportional to the cross-sectional area of the pipe (*A*)**

Darcy's Law

- $Q \alpha h_1 - h_2$
- $Q \alpha \dfrac{1}{L}$
- The flow is obviously proportional to the cross-sectional area of the cylinder (A). When combined with the proportionality constant, K, the result is an expression known as the *Darcy's law*.

Darcy's Law

By introducing a constant of proportionality *K*, *Darcy's experimental results can be summarized as*

$$Q = \frac{K\,A\,(h1 - h2)}{(L)}$$

Where

K = Constant of proportionality commonly known as the *co-efficient of permeability or the hydraulic conductivity*.

Q = Discharge in m^3/day

h_2-h_1= Head loss in metres

***(x1 − x2) =L=* length of medium (flow path) over which head loss occurs measured in direction of flow in metres**

A = cross sectional area of the porous medium though which flow occurs measured at right angle to the flow direction in m^2

Darcy's Law

$\frac{h_1 - h_2}{L}$ is a dimensionless quantity and is often expressed with calculus as dh/dl. Therefore, the Darcy's Law may expressed in more general terms as:

$$Q = - KA \frac{dh}{dl}$$

The above equation is the differential form of Darcy's equation. The quantity dh represents the change in head between two points that are very close together and dl is the small distance between these two points.

Darcy's Law

The dimensionless quantity h1-h2 or dh/dl represents

L

the change in water level elevation in the manometer tubes over the distance through which the change or drop or loss takes place.

Hence this term is referred as gradient or hydraulic gradient and is denoted by i.

Hence, the Darcy's Law in the above equation becomes

$$Q = - KAi$$

CONFINED AQUIFER TESTS

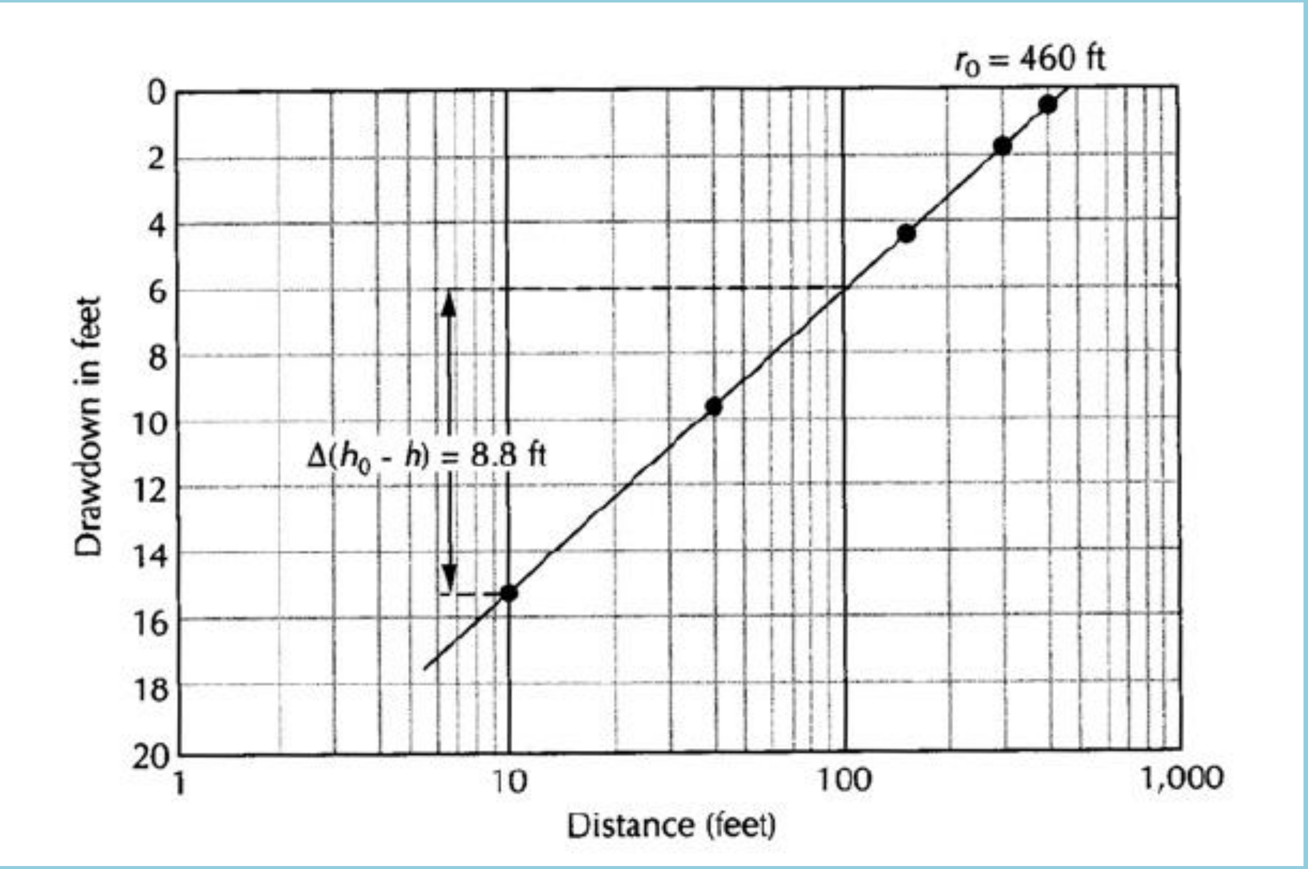
Confined Aquifer & Steady State Flow

Assumptions

- the well fully penetrates the aquifer
- there is a radial flow
- the aquifer is homogeneous and isotropic
- the system prior to pumping is under steady state
- the density and viscosity of water is constant
- the pumping well is of infinitesimal diameter and 100% efficient
- the flow is horizontal and Darcy's law valid
- the potentiometric surface is initially horizontal and changes due to pumping of well
- the aquifer is horizontal, infinite in horizontal extent.
- The aquifer is confined

CONFINED AQUIFER TESTS

Theim's Solution



CONFINED AQUIFER TESTS

Theim's Solution

Procedure I:

$$T = \frac{2.30Q\left(\log(r_2 / r_1)\right)}{2\pi(s_1 - s_2)}$$

Substitute the values of steady state drawdowns and their distances and Q in the equation to get T

CONFINED AQUIFER TESTS

Theim's Solution

Procedure II:

- Plot on a semi logarithmic paper the observed steady-state drawdowns (s) of each piezometer against the distance r. Draw the best fitting line through the plotted points to describe the "distance-drawdown" graph.
- Determine the slope of the line, Δs i.e. the difference of s per log cycle of r, giving r2/r1 = 10 or log r2/r1 = 1. the equation above is reduced to

$$T = \frac{2.30Q}{2\pi\Delta s}$$

CONFINED AQUIFER TESTS

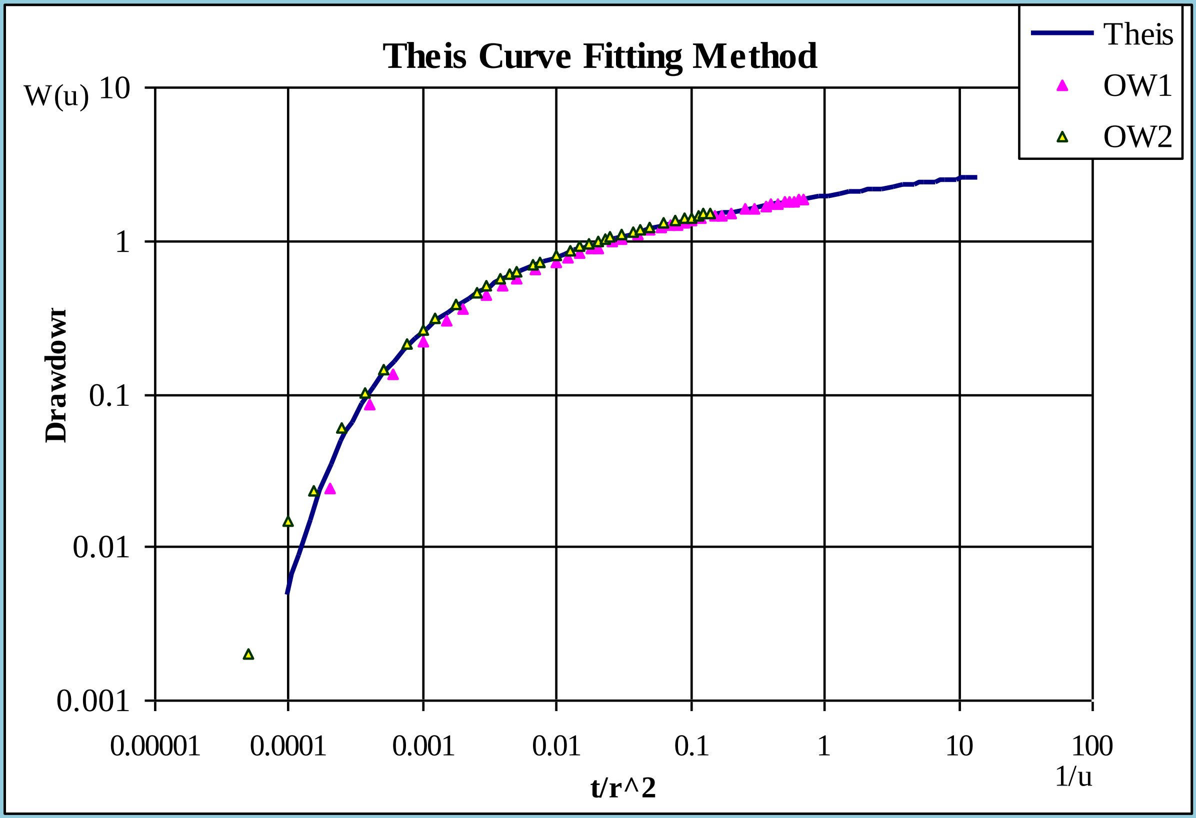
Confined Aquifer - Unsteady State

Theis Method

- The unsteady state drawdown for constant discharge is measured in two or more wells.
- On log-log papers
- W(u) vs 1/u plotted (Standard Curves)
- s vs t plotted
- Two curves superimposed and adjusted to find “Match Point” for one of the curves.
- Values for s, t, W(u) and 1/u found for the Match Point
- Aquifer parameters estimated with the formulae:

CONFINED AQUIFER TESTS

Confined Aquifer - Unsteady State



CONFINED AQUIFER TESTS

Confined Aquifer - Unsteady State

$$KD = \frac{Q}{4\pi s} W(u)$$

$$S = \frac{4KDtu}{r^2}$$

CONFINED AQUIFER TESTS

Confined Aquifer - Unsteady State

The function W(u) is known as known as well-function or Wenzel Function a mathematically expressed as:

$$W(u) = -0.5772 - \ln(u) + u - \frac{u^2}{2\times 2!} + \frac{u^3}{3\times 3!} - .....$$

For Large values of t and small values of r only first two terms are significant.

$$s = \left(\frac{Q}{4\pi T}\right)\left(-0.5772 - \ln(u)\right)$$

CONFINED AQUIFER TESTS

Confined Aquifer - Unsteady State

$$s = \left(\frac{Q}{4\pi T}\right)\left(-0.5772 - \ln(u)\right)$$

$$s = \left(\frac{Q}{4\pi T}\right)\left(\ln(\frac{1}{u}) - \ln(1.781)\right)$$

$$s = \left(\frac{Q}{4\pi T}\right)\left(\ln(\frac{1}{u}) - \ln(1.781)\right)$$

$$s = \left(\frac{Q}{4\pi T}\right)\left(\ln(\frac{1}{u}) - \ln(1.781)\right)$$

$$s = \left(\frac{2.30Q}{4\pi T}\right)\left(\log(\frac{0.5615 \times 4Tt}{r^2 S})\right)$$

$$s = \left(\frac{2.30Q}{4\pi T}\right)\left(\log(\frac{2.25Tt}{r^2 S})\right)$$

CONFINED AQUIFER TESTS

Confined Aquifer - Unsteady State

$$s = \left(\frac{2.30Q}{4\pi T}\right)\left(\log(\frac{2.25Tt}{r^2 S})\right)$$

$$s = \left(\frac{2.30Q}{4\pi T}\right)\left(\log(\frac{2.25T}{r^2 S})\right) + \left(\frac{2.30Q}{4\pi T}\right)(\log(t))$$

This is a straight line equation s Vs Log(t)

At the intercept s=0 hence

$$0 = \left(\frac{2.30Q}{4\pi T}\right)\left(\log(\frac{2.25Tt_0}{r^2 S})\right)$$

As

$$\left(\frac{2.30Q}{4\pi T}\right) \textit{cannot be zero}$$

$$\left(\log(\frac{2.25Tt_0}{r^2 S})\right) = 0$$

$$\frac{2.25Tt_0}{r^2 S} = 1$$

$$S = \frac{2.25Tt_0}{r^2}$$

CONFINED AQUIFER TESTS

Confined Aquifer - Unsteady State

$$s = \left(\frac{2.30Q}{4\pi T}\right)\left(\log(\frac{2.25T}{r^2 S})\right) + \left(\frac{2.30Q}{4\pi T}\right)(\log(t))$$

The derivation with respect to log(t) is

$$\Delta s = \left(\frac{2.30Q}{4\pi T}\right)\Delta(\log(t))$$

$$T = \left(\frac{2.30Q}{4\pi \Delta s}\right)\Delta(\log(t))$$

slope per logcycle

$$T = \left(\frac{2.30Q}{4\pi \Delta s}\right)$$

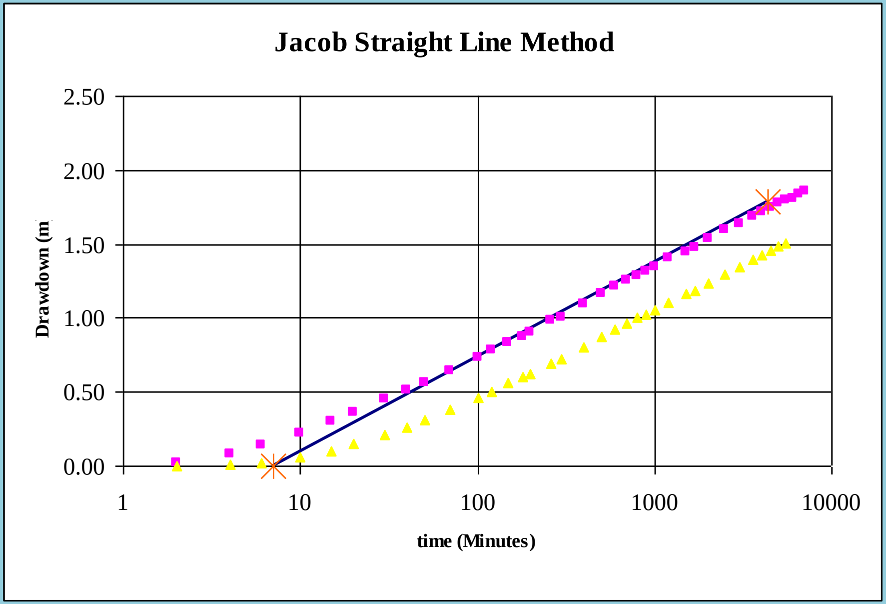
CONFINED AQUIFER TESTS

- **Jacob Straight Line Method**

For combined estimation of parameters for all the OW same method is used by plotting s vs t (log)

$$T = \left(\frac{2.30Q}{4\pi \Delta s} \right)$$

$$S = \frac{2.25Tt_0}{r^2}$$



CONFINED AQUIFER TESTS

Recovery Method

Theis Recovery Method

- According to Theis (1935), the residual drawdown after a pumping test with a constant discharge is

$$s' = \frac{Q}{4\pi KD}\left(W(u) - W(u')\right)$$

where $u = \frac{r^2 S}{4KDt}$ $\quad u' = \frac{r^2 S}{4KDt'}$

- When u and u’ are sufficiently small. (< 0.01), the above equation can be approximated by

$$s' = \frac{Q}{4\pi KD}\left(\ln\frac{4KDt}{r^2 S} - \ln\frac{4KDt'}{r^2 S}\right)$$

CONFINED AQUIFER TESTS

Theis Recovery Method

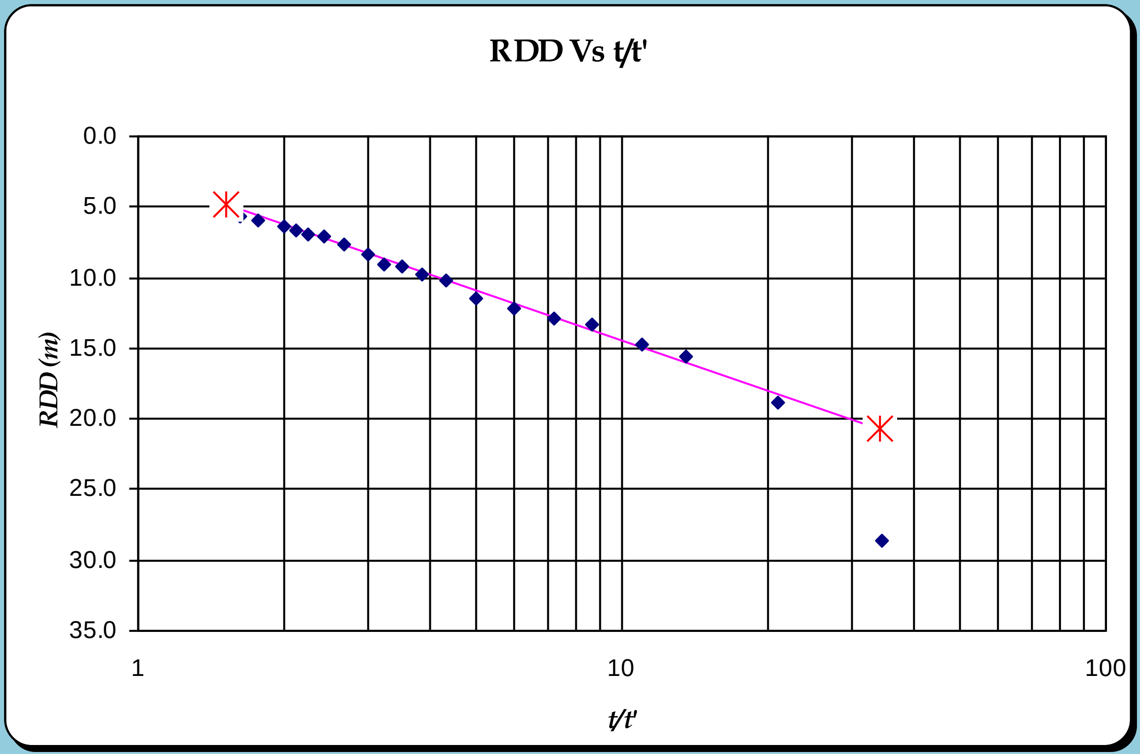
- For each observed value of s‘, calculate the corresponding value of t/t‘;
- For one of the piezometers, plot s’ versus t/t’ on semi-log paper (t/t’ on the log scale)
- Fit a straight line through the plotted points;
- Determine the slope of the straight line, i.e. the residual drawdown difference Δs
- Substitute the known values of Q and Δs’ into following equation and calculate KD.

$$KD = \frac{2.30Q}{4\pi\Delta s'}$$

- It is suitable for tests in single pumping wells

CONFINED AQUIFER TESTS

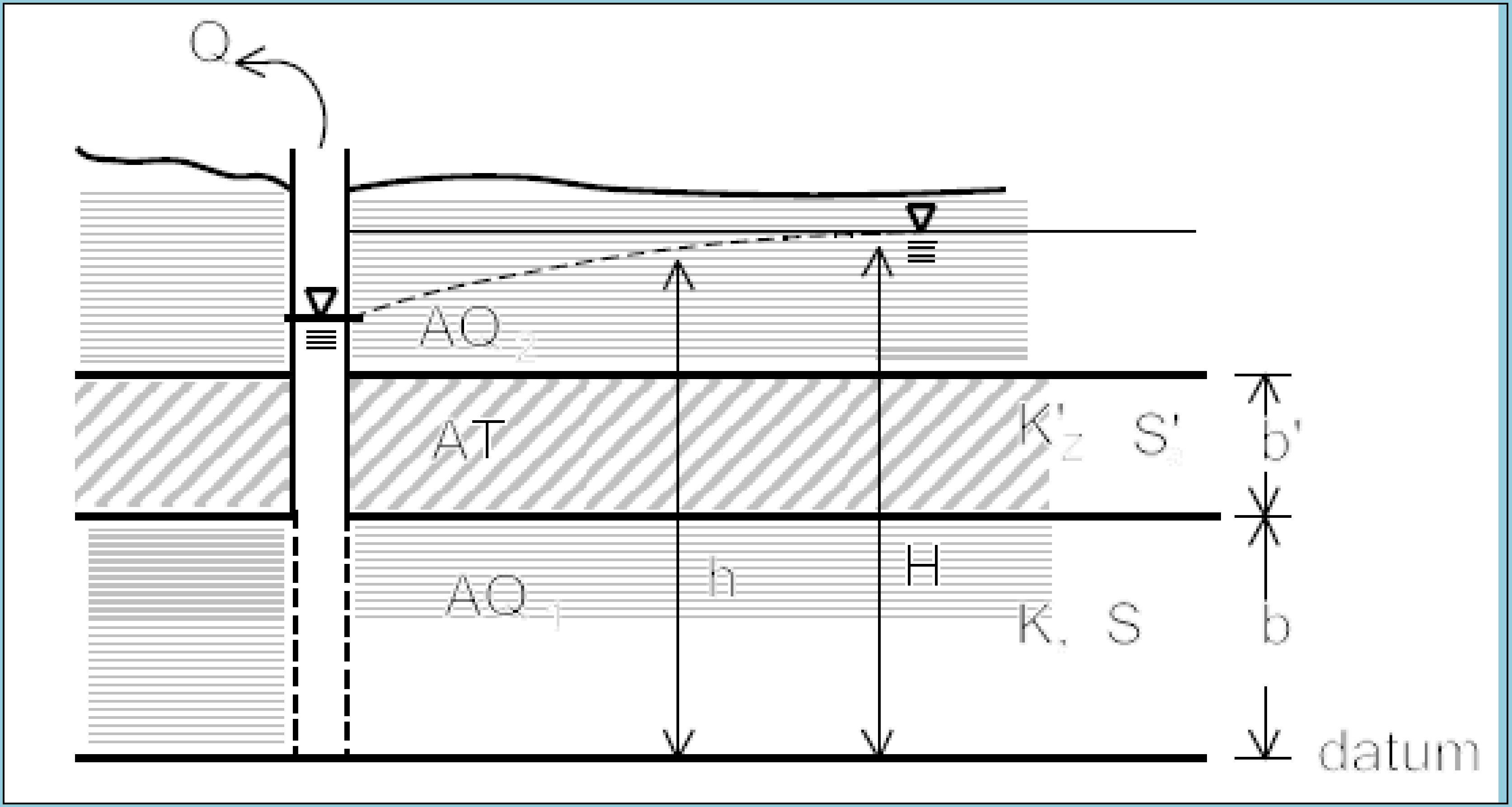
Theis Recovery Method



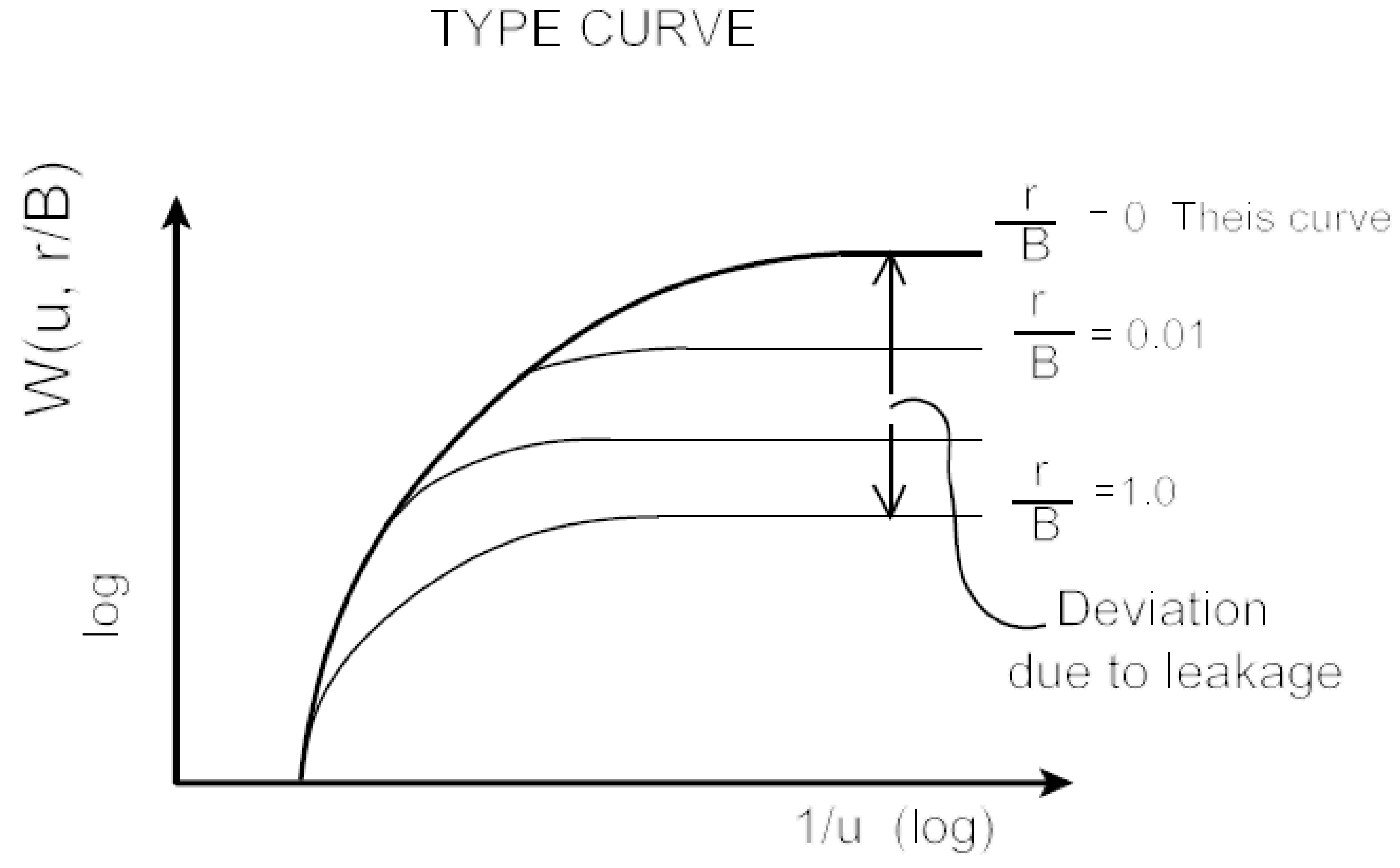
Date of Test:	17/12/02	
SWL:	13.07	m
Q:	2.00	lps
Pumping Time:	100.00	Min
$\triangle$ s	11.77	m
T	2.69	m^2/day

SEMI-CONFINED AQUIFER TESTS

SEMI-CONFINED AQUIFER TESTS



SEMI-CONFINED AQUIFER TESTS



SEMI-CONFINED AQUIFER TESTS

Two Dimensional Flow Equation

$$T\left(\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}\right)+q_z=-S\frac{\partial h}{\partial t}$$

where q_z represents the inflow from the aquitard above the aquifer and it is defined as

$$q_z = K'_z \frac{(H-h)}{b'}$$

K'_Z and b' are the vertical hydraulic conductivity and the thickness of the aquitard

$$\left(\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}\right)+K'_z\frac{(H-h)}{Tb'}=-\frac{S}{T}\frac{\partial h}{\partial t}$$

SEMI-CONFINED AQUIFER TESTS

Taking polar coordinates and $(H - h) = s$ as drawdown, the equation can be expressed as

$$\left(\frac{\partial^2 s}{\partial r^2}\right) + \frac{1}{r}\left(\frac{\partial s}{\partial r}\right) - \frac{s}{B^2} = \frac{S}{T}\frac{\partial s}{\partial t}$$

$$\left(\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2}\right) + K'_z \frac{(H-h)}{Tb'} = -\frac{S}{T}\frac{\partial h}{\partial t}$$

$$\text{Leakage Factor} = B = \sqrt{Tb' / K'_Z}$$

b' / K'_Z is called Hydraulic Resistance (c)

Leakage factor is expresses in meter. Hence large leakage factor means less leakage through the aquitard.

SEMI-CONFINED AQUIFER TESTS

The assumptions and conditions for the methods used for analysis of pumping test data in leaky aquifers are :

- The aquifer is leaky
- The aquifer and the aquitard have a seemingly infinite areal extent
- The aquifer and the aquitard are homogeneous, isotropic and of uniform thickness over the area influenced by the test.
- Prior to pumping, the piezometric surface and the water table are horizontal over the area of influence.
- The aquifer is pumped at a constant discharge rate
- The well penetrates the entire thickness of the aquifer and receives water by horizontal flow.

SEMI-CONFINED AQUIFER TESTS

- The flow in the aquitard is vertical
- The drawdown in the upper aquifer and the aquitard is negligible

Additional assumptions for the unsteady state flow are:

- The water removed from storage in the aquifer and the water supplied by leakage from the aquitard is discharged instantaneously with decline of head
- The diameter of the wells is very small and storage within the well can be neglected.

SEMI-CONFINED AQUIFER TESTS

Steady State	De Glee Method Hantush Jacob Method
Unsteady State	Hantush - 1955 (Walton) Method Hantush Inflection Point Method Hantush - 1960 Method

SEMI-CONFINED AQUIFER TESTS

Steady State

Steady State	De Glee Method Hantush Jacob Method

SEMI-CONFINED AQUIFER TESTS

De Glee's Method

De Glee derived a formula for steady state flow to an aquifer with leakage from an aquitard proportional to the hydraulic gradient across the aquitard.

$$\boxed{s_m = \frac{Q}{2\pi KD} K_0\left(\frac{r}{L}\right)}$$

where

s_m = steady state (stabilised) drawdown (m) in a Piezometer at a distance of r (m) from the well

Q = discharge of the well (m^3/day)

L= $\boxed{L = \sqrt{KDc}}$ - leakage factor (m)

c = D'/K' - hydraulic resistance of the aquitard (days)

D'= saturated thickness of the aquitard

K'= hydraulic conductivity of the aquitard for vertical flow (m/day)

K_0(x)= modified Bessel Function of the second kind and zero order.

De Glee's Method

Procedure

- On one log-log paper $K_0(x)$ vs r/L to prepare master curve
- On another log-log paper s vs r is plotted
- Two curves superimposed and adjusted to find a "Match Point".
- Values for s, r, $k_0(r/L)$ and r/L are read for the Match point
- Parameters are estimated with the formulae:

$$\boxed{KD = \frac{Q}{2\pi s_m} K_0\left(\frac{r}{L}\right)}$$

$$\boxed{c = \frac{L^2}{KD} = \frac{1}{(r/L)^2} \times \frac{r^2}{KD}}$$

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Hantush-Jacob Method

Hantush observed that if r/L is small ($r/L \leq 0.05$), the above equation can be approximated as:

$$s_m \approx \frac{2.30Q}{2\pi KD}\left(\log 1.12\frac{r}{L}\right)$$

A plot of s_m against r on semi-log paper, with r on logarithmic scale would show a straight line for small values of r/L.

The slope of the straight line portion of the curve, i.e., the drawdown difference Δs_m per log cycle of r, is expressed as:

$$\Delta s_m = \frac{2.30Q}{2\pi KD}$$

$$KD = \frac{2.30Q}{2\pi \Delta s_m}$$

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The straight-line portion of the curve is extended to intercept the r axis where the drawdown is zero. At the interception point, $s_m = 0$ and $r = r_0$ and thus the above equation reduces to:

$$1.12\frac{L}{r_0} = \frac{1.12}{r_0}\sqrt{KDc} = 1$$

$$c = \frac{(r_0/1.12)^2}{KD}$$

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Unsteady State

Unsteady State	Hantush - 1955 (Walton) Method Hantush Inflection Point Method Hantush - 1960 Method

Walton's Method

The drawdown in a leaky aquifer under unsteady state condition and with negligible aquitard storage can be described by the following formula given by Hantush & Jacob .

$$\boxed{s = \frac{Q}{4\pi KD}\int_u^{\infty}\frac{1}{y}\exp\left(-y-\frac{r^2}{4L^2 y}\right)dy}$$

$$\boxed{s = \frac{Q}{4\pi KD}W(u, r/L)}$$

$$\boxed{u = \frac{r^2 S}{4KDt}}$$

This equation is similar to the Theis but in the well function there are two parameters in the integral, u and r/L. The method is similar to Theis's curve fitting method, but, instead of one curve, there is a family of type curves for different values of r/L.

Walton's Method

Procedure:

- Plot on a double log paper values of W(u,r/L) versus 1/u for different values of r/L to get a Walton Family Type Curves.
- On another transparent sheet of double log paper of the same scale, plot the values of s versus t for one of the observation wells.
- Superimpose the field data plot on the type curves and adjust the field plot till it matches well with one of the Walton type curves.
- Select an arbitrary point **"A"** on the two graphs and note the values of W(u,r/L), 1/u, s, t, and r/L.
- Substitute these values in the following equations and calculate the values of KD and S.

$$KD = \frac{Q}{4\pi s} W(u, r/L) \qquad S = \frac{4KDt}{r^2} . u$$

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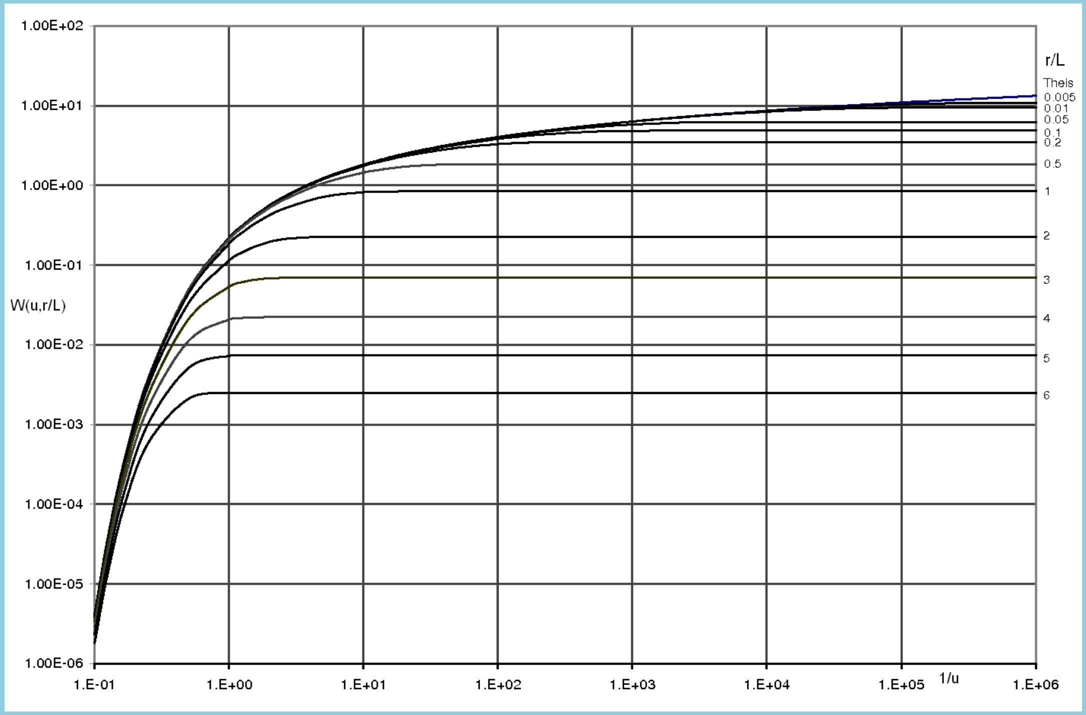
Hantush – 1955 (Walton) Method

- Calculate the value of L from the noted value of r/L and known value of r and
- Calculate the value of c as $c=L^2/KD$



Hantush Inflection Point Method

Procedure

- Plot the value of drawdown (s) for on observation well against corresponding value of t on a semi-log paper (t on log scale)
- Record the observed/extrapolated value of steady state drawdown s_m
- Calculate the value of s_p as $s_p = 0.5 s_m$ and locate the "inflection point" p on the field curve.
- Record the time value t_p at the inflection point.
- Draw a tangent to the curve at the inflection point and determine the slope of the curve Δs_p (the drawdown difference for one log cycle of time).:

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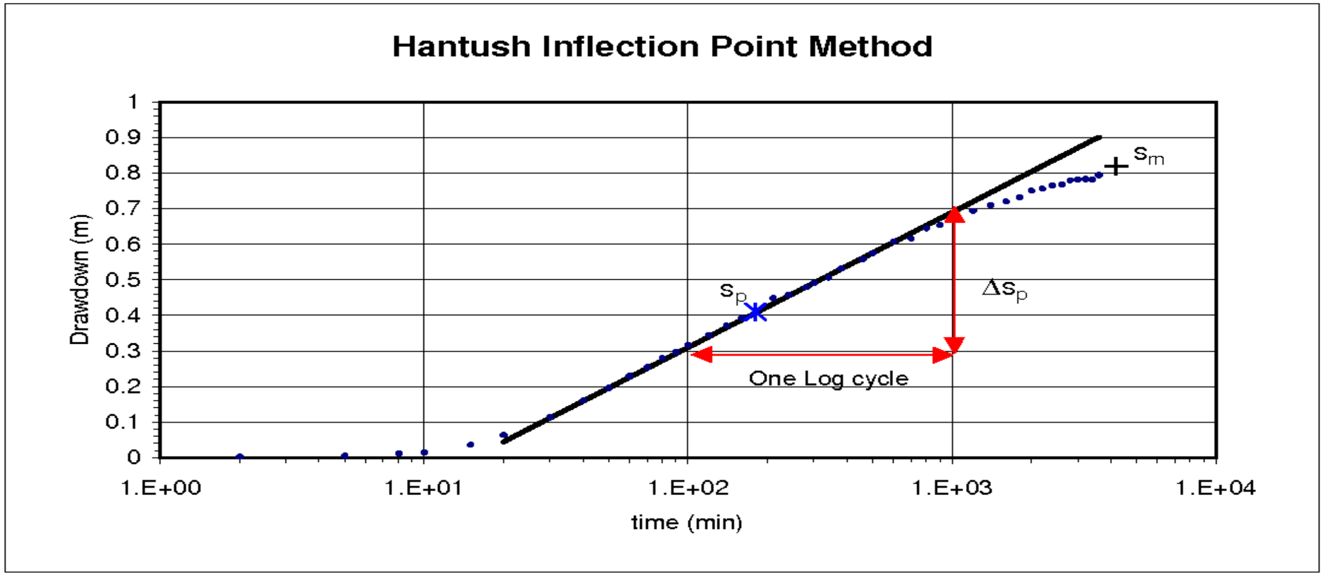
Hantush Inflection Point Method



Site: **Dakoha**

Well:	**OW**		**Match Point**			**Parameters**		
Q=	5077	m3/d	s_m=	0.82	min	$e^{r/L}K0(r/L)$	2.481579	
r=	200	m	s_p=	0.41	min	r/L	0.13	
D=	86	m	t_p=	180	min	$e^{r/L}$	0.878	
D'=	17	m	t_p=	0.125	day	L	1538	m
			Δs_p	0.38	-	KD	2147	m2/day
						S	1.74E-03	(-)
						c	1102	days
						K'	0.02	m/day

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Hantush Inflection Point Method

Procedure

$$2.30\frac{s_p}{\Delta s_p} = e^{-r/L}K_0(r/L) = f(r/L)$$

- Calculate the value of f(r/L) from the above equation by substituting values of Δs_p and s_p. From standard table, determine the value of r/L for computed value of f(r/L).
- Calculate value for L from values of r and r/L
- Record the value of $e^{-r/L}$ from the standard table.
- Calculate the value of KD from the known and/or calculated values of Q, Δs_p and $e^{-r/L}$ using the equation:

$$KD = \frac{2.30Q}{4\pi\Delta s_p}e^{-r/L}$$

Hantush Inflection Point Method

Procedure

$$u_p = \frac{r^2 S}{4\pi KDt_p} = \frac{r}{2L}$$

- S can be calculated from the values of KD, t_p, r and r/L.
- Calculate the value of c from $c = L^2/KD$.

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Hantush -1960 Method

Procedure

- On a double log paper, plot the values of $W(u,\beta)$ versus values of $1/u$ for different values of β to get the family of type curves.
- Plot on another sheet of transparent log paper of the same scale the values of s versus t.
- Superimpose this field data on plot type curve keeping the $W(u,\beta)$ axis parallel to the s axis and $1/u$ axis parallel to the t axis.
- Adjust till the field curve matches with one of the type curves.
- Select a match point **"A"** on both the graphs and record the values of $W(u,\beta)$, $1/u$, s, t and β for this match point.

SEMI-CONFINED AQUIFER TESTS

Hantush -1960 Method

Procedure

- Substitute these values in the equations given below and calculate the values of KD and S.

$$KD = \frac{Q}{4\pi s} W(u,\beta)$$

$$S = \frac{4KDt}{r^2} u$$

- From the values of β, KD, S, r, and D', calculate the value of K'S'.

$$\frac{K'S'}{D'} = \beta^2 (4/r)^2 KDS$$

Thank You