Groundwater

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- Aquifer Properties
- Aquifer Properties
- Darcy’s Law
- Coefficient of Permeability
- Transmissibility
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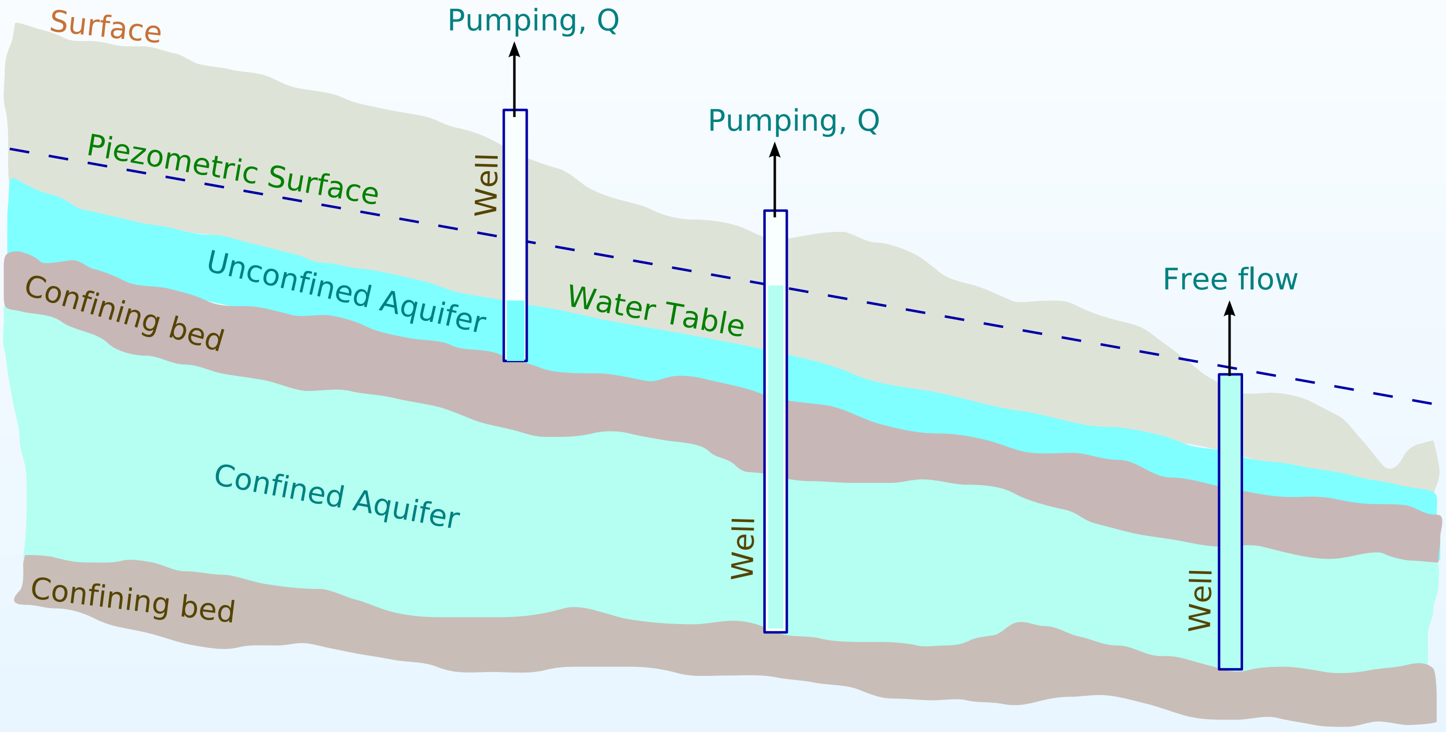

Confined and Unconfined Aquifers

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Geological Formations

On the basis of availability of water from the saturated zone below the ground surface, geological formations are classified into four categories:

1. Aquifer: A saturated formation of earth material that not only stores water but also yields it in sufficient quantity. (*Deposits of sand and gravel*)

2. Aquitard: A formation through which only seepage is possible and thus the yield is insignificant compared to an aquifer. (*Sandy clay*)

3. Aquiclude: A geological formation that is essentially impermeable to the flow of water. It may contain large amount of water due to its high porosity. (*Clay*)

4. Aquifuge: A geological formation, that is neither porous or permeable. There are no interconnected openings and hence it cannot tranmit water. (*Compact rock*)

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Aquifer Properties

Porosity

The amount of pore space per unit volume of the aquifer material is the *porosity*, η

$$\eta = \frac{V_v}{V_0}$$

where η = porosity, V_v = volume of voids and V_0 = volume of the porous medium.

In quantitative terms, porosity greater than 20% is considered as large, between 5 to 20% as medium and less than 5% as small.

Specific Yield

The actual volume of water that can be extracted by the force of gravity from a unit volume of aquifer material is known as *Specific Yield*, S_y.

The fraction of water held back in the aquifer is known as *Specific Retention*, S_r.

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Aquifer Properties

Porosity and Specific Yield of Selected Formations

Formation	Porosity (%)	Specific Yield (%)
Clay	45-55	1-10
Sand	35-40	10-30
Gravel	30-40	15-30
Sand stone	10-20	5-15
Shale	1-10	0.5-5
Lime stone	1-10	0.5-5

Although both clay and sand have similar high porosity, specific yield of clay is comparatively much less.

Darcy's Law

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Velocity of flow through a porous medium is directly proportional to the available hydraulic gradient.

$$V = K\,i$$

V = Apparent velocity of seepage flow; K = a coefficient, called coefficient of permeability; $i = -\dfrac{dh}{ds}$ = hydraulic gradient, h = piezometric head and s = the distance measured in the general flow direction. The negative sign indicates that the piezometric head drops in the direction of flow.

It may be noted that the apparent velocity V is not the actual velocity of flow through pores. Due to the irregular pore geometry, the actual velocity of flow varies from point to point and the *bulk pore velocity*, v_a, which represents the actual speed of travel of water in porous media is expressed as

$$v_a = \frac{V}{n}$$

where n = porosity.

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Coefficient of Permeability

The coefficient of permeability, also called the *hydraulic conductivity*, reflects the combined effects of the porous medium and fluid properties. It may be expressed as

$$K = C\, d_m^2\, \frac{\gamma}{\mu}$$

where d_m = mean particle size of the porous medium, $\gamma = \rho g$ = unit weight of fluid, ρ = density of the fluid, g = acceleration due to gravity, μ = dynamic viscosity of the fluid and C = a shape factor that depends on porosity, packing, shape of grains and grain-size distribution of the porous medium.

The coefficient of permeability is often considered in two components: one that reflects the medium properties and the other that reflects fluid properties.

$$K = C\, d_m^2\, \frac{\gamma}{\mu} = K_0\, \frac{\gamma}{\mu} = K_0\, \frac{g}{\nu}$$

where $K_0 = C\, d_m^2$ is called specific or *intrinsic permeability*. It is a fucntion of the medium only. It has the dimension of $[L_0]$.

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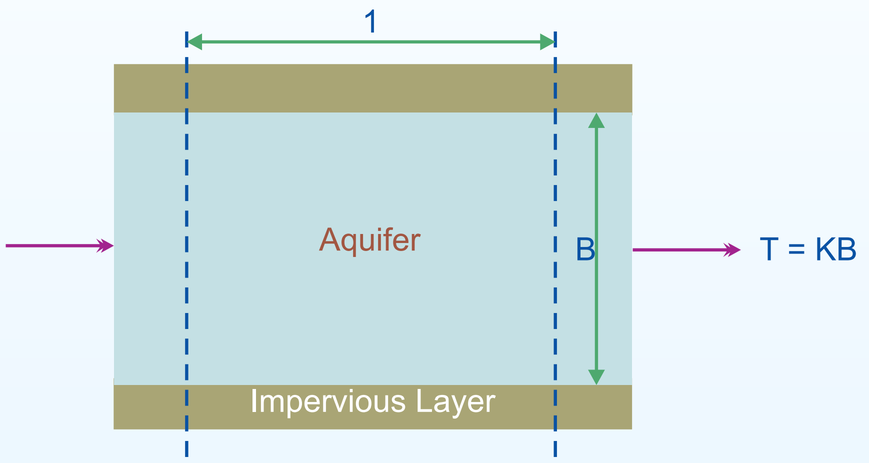
Transmissibility

Consider an aquifer of *unit width* and thickness B (depth of the fully saturated zone).



The discharge through this aquifer under a *unit hydraulic gradient* is $T = KB$. This discharge is called *transmissibility*. It has the dimension of $[L^2/T]$. Typical units are m2/s or litres per day/metre width.

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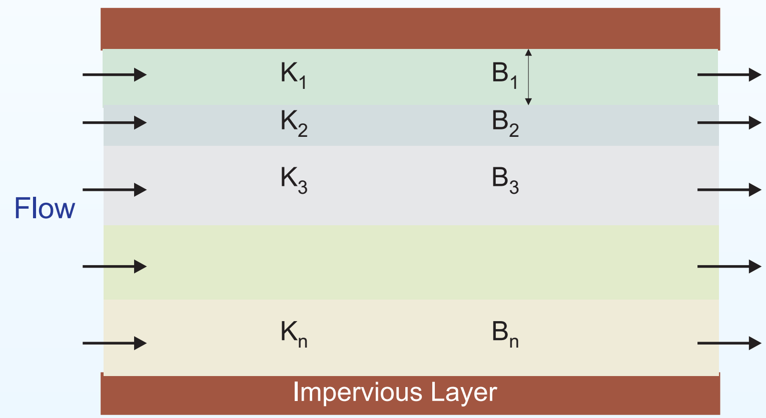
Stratification

Sometimes the aquifers may be stratified, with different permeabilities in each strata. Stratification may be *parallel* or *normal* to the flow.

Parallel Stratification



When the flow is parallel to the stratification, the equivalent permeability K_e of the entire aquifer of thickness B is

$$K_e = \frac{\sum K_i B_i}{\sum B_i}$$

and the transmissivity is $T = K_e \sum B_i = \sum K_i B_i$

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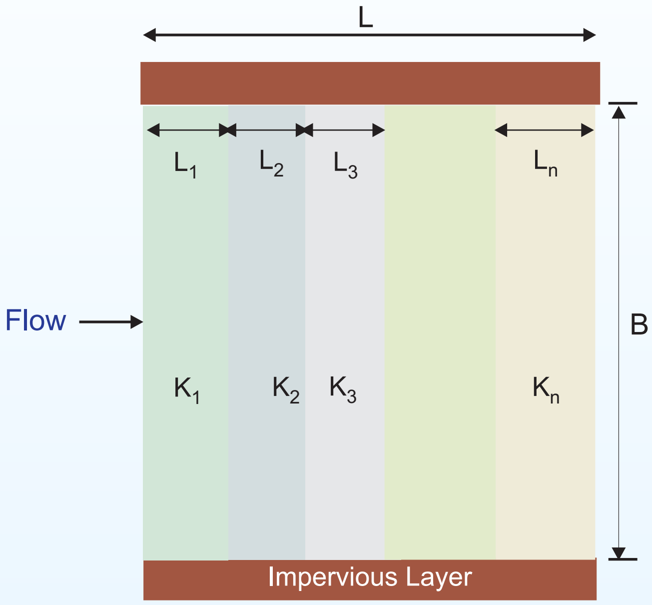
Normal Stratification



When the flow is normal to the stratification, the equivalent permeability K_e of the aquifer of length $L = \sum L_i$ is

$$K_e = \frac{\sum L_i}{\sum (L_i/K_i)}$$

The transmissivity is $T = K_e \cdot B$

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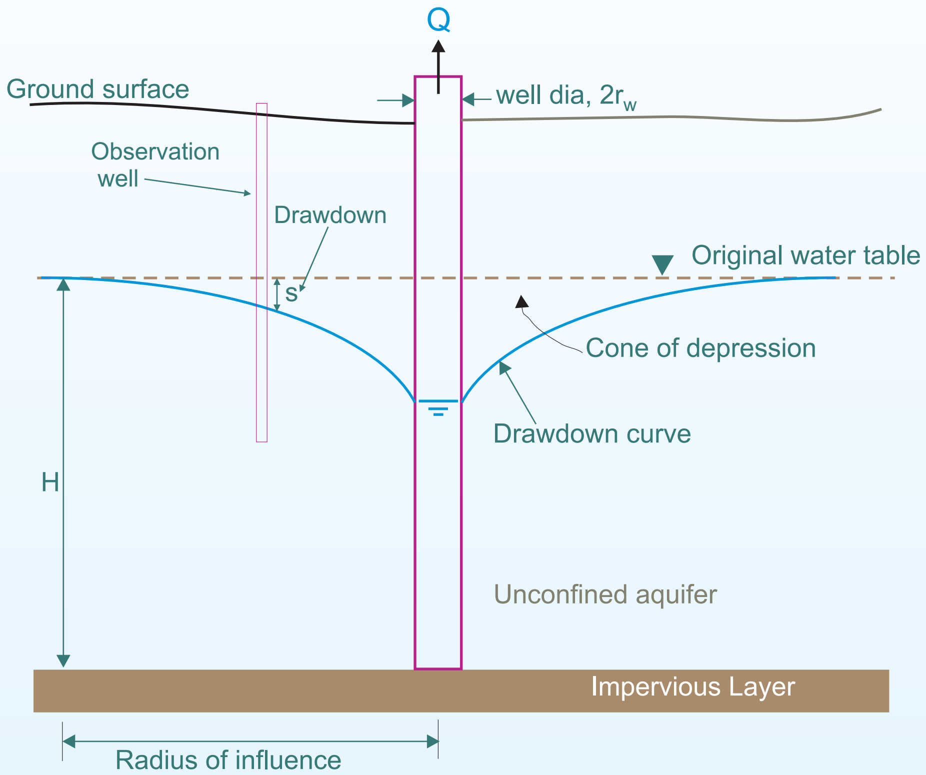
Introduction

Wells form the most important mode of groundwater extraction from an aquifer. These find extensive in water supply and irrigation engineering practices.



Consider the water in an unconfined aquifer being pumped at a constant rate from a well. Prior to pumping, water level in the indicates the static water table. Lower of this water table takes place on pumping.

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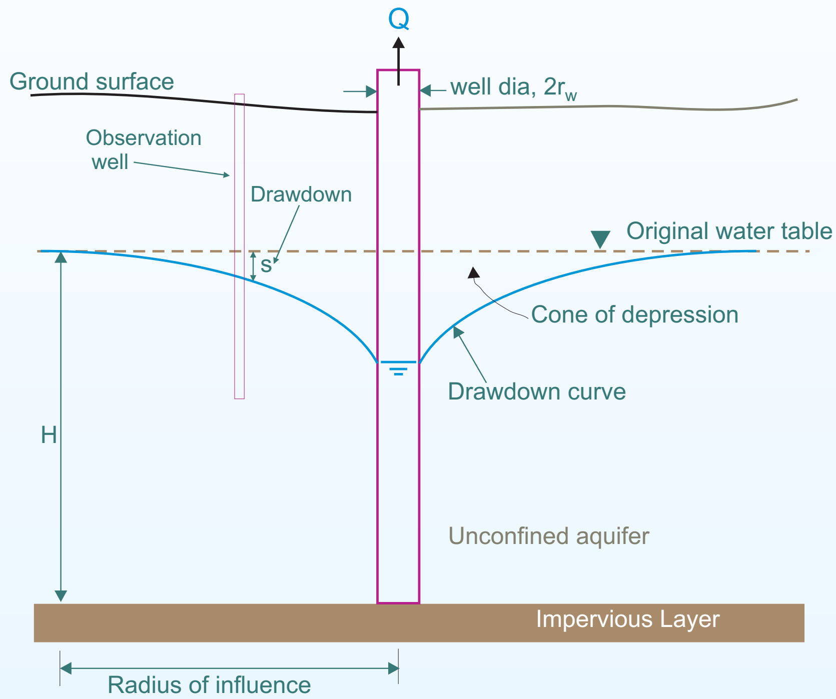

If the aquifer is homogeneous and isotropic and the water table is horizontal initially, then due to the radial flow into the well through the aquifer, the water table assumes a conical shape called *cone of depression*. The areal extent of the cone of depression is called *area of influence* and its radial extent *radius of influence*.

The drop in the water table elevation at any point from its previous static level is called *drawdown*.

State of Flow

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At constant rate of pumping, the drawdown curve develops gradually with time due to the withdrawal of water from storage. This phase is called *unsteady flow*, as the water table at a given location near the well changes with time.

On prolonged pumping, an equilibrium state is reached between the rate of pumping and the rate of inflow of groundwater from outer edges of the zone of influence. The drawdown surface attains a constant position with respect to time when the well is known to operate under *steady flow* conditions.

As soon as the pumping is stopped, the depleted storage in the cone of depression is made good by groundwater inflow into the zone of influence. There is a gradual accumulation of storage till the original static level is reached. This stage is called *recovery* and is an unsteady phenomenon.

Similar changes take place to a well in *confined* aquifer, but the difference is that, it is the piezometric surface instead of water table that undergoes drawdown with the development of cone of depression.

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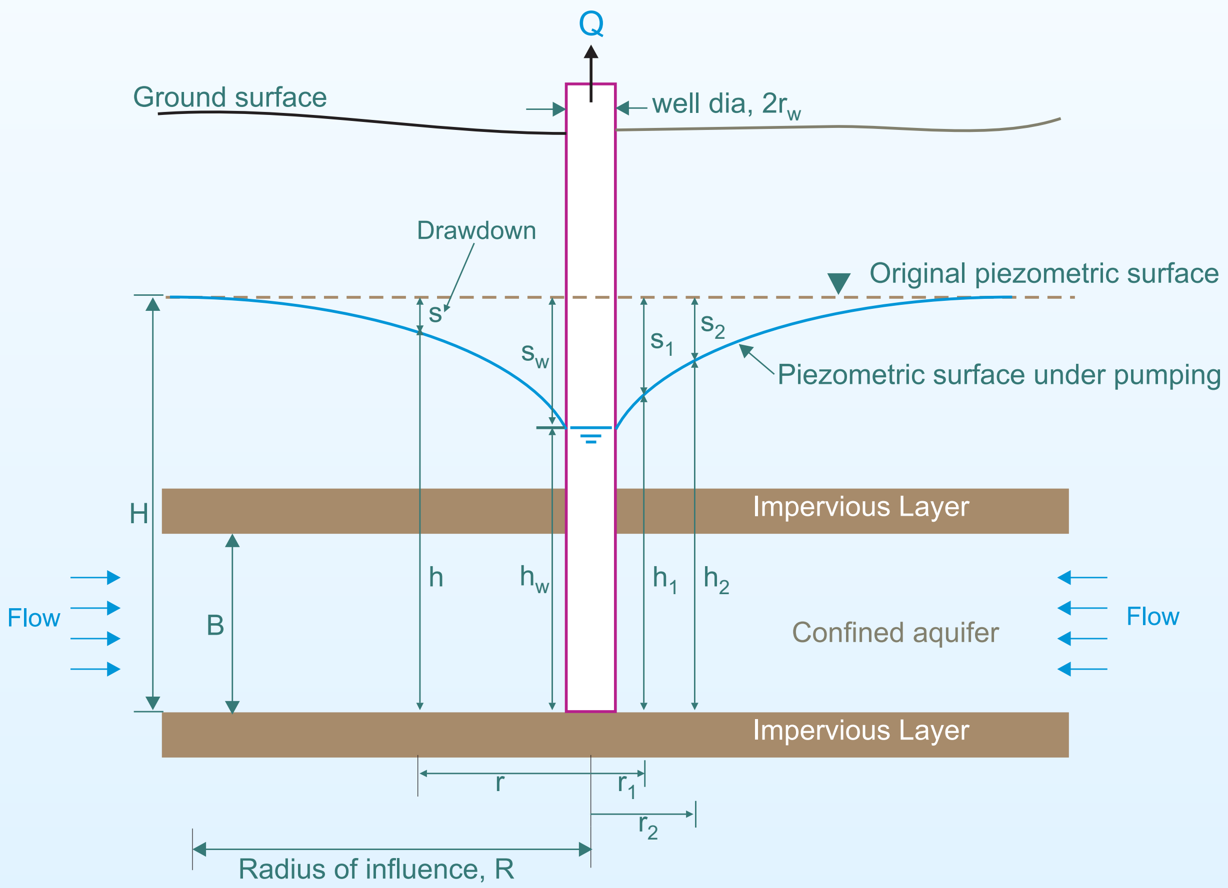
Confined Flow

The figure below shows a well completely penetrating a horizontal confined aquifer of thickness B. Consider the well to be discharging a steady flow, Q. The original piezometric head was H and the drawdown due to pumping is indicated in the figure. The piezometric head the the well is h_w and the drawdown is s_w.



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Confined Flow

At a radial distance r from the well, if h is the piezometric head, then velocity of flow is

$$V_r = K\frac{dh}{dr}$$

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Confined Flow

At a radial distance r from the well, if h is the piezometric head, then velocity of flow is

$$V_r = K\frac{dh}{dr}$$

The cylindrical surface through which this velocity occurs is $2\pi rB$. Then the discharge is

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$$V_r = K\frac{dh}{dr}$$

The cylindrical surface through which this velocity occurs is $2\pi rB$. Then the discharge is

$$\begin{aligned} Q &= (2\pi rB)\left(K\frac{dh}{dr}\right) \\ \frac{Q}{2\pi KB}\frac{dr}{r} &= dh \end{aligned}$$

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$$\frac{Q}{2\pi KB}\frac{dr}{r} = dh$$

Integrating between limits r_1 and r_2 with piezometric heads $h_1$and h_2,

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Integrating between limits r_1 and r_2 with piezometric heads $h_1$and h_2,

$$\begin{aligned} \frac{Q}{2\pi KB}\ln\frac{r_2}{r_2} &= (h_2 - h_1) \\ Q &= \frac{2\pi KB(h_2 - h_1)}{ln(r_1/r_2)} \end{aligned}$$

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If the drawdown s_1 and s_2 at the observation wells are known, then

$$Q = \frac{2\pi T(s_1 - s_2)}{\ln(r_2/r_1)}$$

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If the drawdown s_1 and s_2 at the observation wells are known, then

$$Q = \frac{2\pi T(s_1 - s_2)}{\ln(r_2/r_1)}$$

because, $s_1 = H - h_1$; $s_2 = h - h_2$ and $KB = T$

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because, $s_1 = H - h_1$; $s_2 = h - h_2$ and $KB = T$

Further, at the edge of the zone of influence, $s = 0$, $r_2 = R$ and $h_2 = H$. Again at the well $s_1 = s_w$, $r_1 = r_w$ and $h_1 = h_w$. Then

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$$Q = \frac{2\pi T s_w}{\ln(R/r_w)}$$

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$$Q = \frac{2\pi T s_w}{\ln(R/r_w)}$$

These two equations can be used to estimate T and hence, K , from pumping tests.

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Example

A 30 cm diameter well completely penetrates a confined aquifer of permeability 45 m/day. The length of the strainer (B) is 20m. Under steady state of pumping the drawdowns at the well was found to be 3.0m and the radius of influence was 300m. Calculate the discharge.

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Solution

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Solution

Here, radius of the well, r_w = 0.15m, radius of influence, R = 300m, drawdown at well, s_w = 3.0m, strainer width, B = 20m, coefficient of permeability, K = 45m/day = 5.208 x 10^{-4} m/s, $T = KB =$ 10.416 x 10^{-3} m^2/s.

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Then discharge,

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Then discharge,

$$Q = \frac{2\pi T s_w}{\ln(R/r_w)} = \frac{2\pi \times 10.416 \times 10^{-3} \times 3}{\ln(300/0.15)} = 0.02583\, m^3/s = 1550 lpm$$

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Prob. 1

For the well in the previous example, calculate the discharge a) if the well diameter is 45cm and all other data remain same, (b) if the drawdown is increased to 4.5cm and all other data remain unchanged.

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For the well in the previous example, calculate the discharge a) if the well diameter is 45cm and all other data remain same, (b) if the drawdown is increased to 4.5cm and all other data remain unchanged.

Ans: (a)

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Prob. 1

For the well in the previous example, calculate the discharge a) if the well diameter is 45cm and all other data remain same, (b) if the drawdown is increased to 4.5cm and all other data remain unchanged.

Ans: (a) 1637 lpm;

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Problem

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For the well in the previous example, calculate the discharge a) if the well diameter is 45cm and all other data remain same, (b) if the drawdown is increased to 4.5cm and all other data remain unchanged.

Ans: (a) 1637 lpm; (b) 2325 lpm

Prob. 2

The discharge from a fully penetrating well operating under steady state in a confined aquifer of 35m thickness is 3000 lpm. Values of drawdown at two observation wells 12m and 120m away from the well are 3.0m and 0.30m respectively. Determine the permeability of the aquifer.

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The discharge from a fully penetrating well operating under steady state in a confined aquifer of 35m thickness is 3000 lpm. Values of drawdown at two observation wells 12m and 120m away from the well are 3.0m and 0.30m respectively. Determine the permeability of the aquifer.

Ans:

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Ans: 16.75 m/day.

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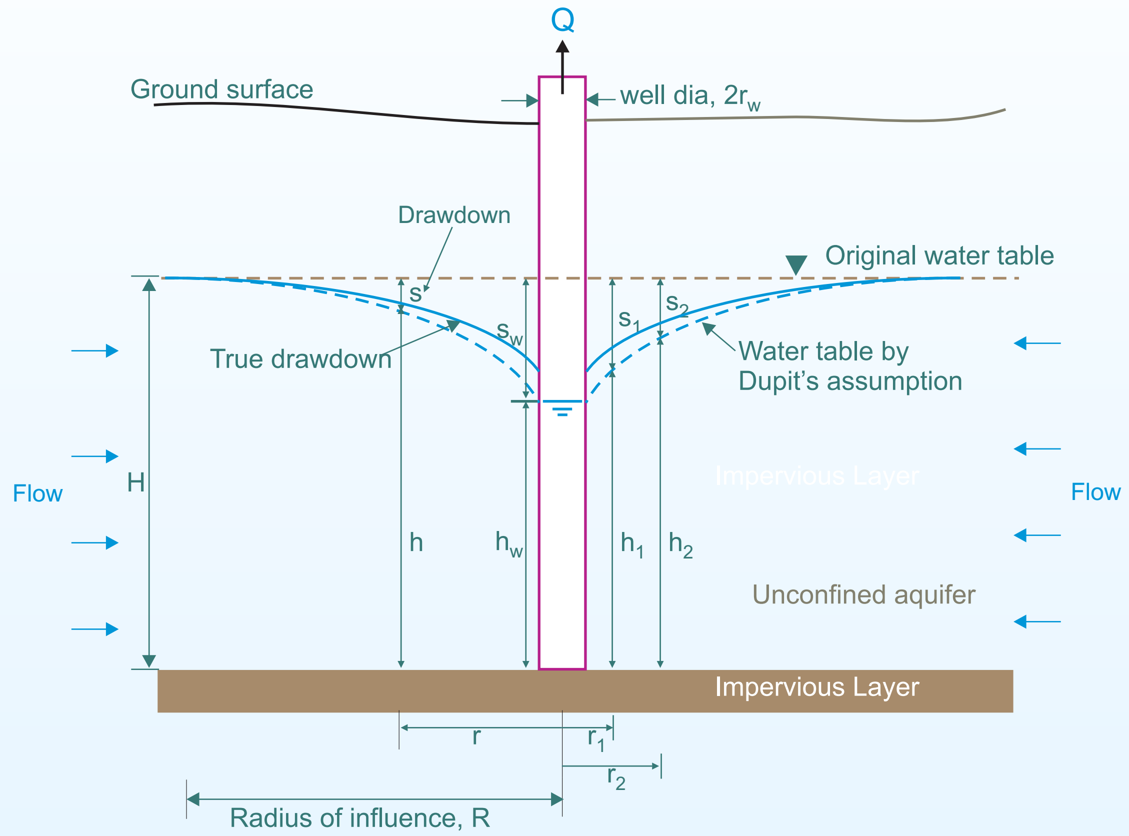
Unconfined Flow



Consider a steady flow from a well completely penetrating an unconfined aquifer. In this case, because of the presence of a curved free surface, the streamlines are not strictly radial straight lines. To obtain a simple solution, *Dupit's Assumptions* are made.

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Dupit’s Assumptions

1. For small inclinations of the free surface, the streamlines can be assumed to be horizontal and equipotentials are thus vertical.

2. The hydraulic gradient is equal to the slope of the free surface and does not vary with depth. This assumption is satisfacory in most of the flow regions except in immediate neighbourhood of the well.

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Consider the well of radius r_w completely penetrating an extensive unconfined horizontal aquifer, as shown in the previous figure. The well is pumping a discharge Q. At any distance r, the velocity of radial flow into the well is

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Consider the well of radius r_w completely penetrating an extensive unconfined horizontal aquifer, as shown in the previous figure. The well is pumping a discharge Q. At any distance r, the velocity of radial flow into the well is

$$V_r = K\frac{dh}{dr}$$

where h is the height of the water table above the aquifer bed at that location. For steady flow, from continuity equation,

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$$Q = (2\pi rh)V_r = 2\pi rKh\frac{dh}{dr}$$

$$\frac{Q}{2\pi K}\frac{dr}{r} = hdh$$

Integrating between limits r_1 and r_2 where the water table depths are h_1 and h_2 and rearranging,

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$$Q = \frac{\pi K(h_2^2 - h_1^2)}{\ln(r_2/r_1)}$$

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At the edge of the zone of influence of radius R, saturated thickness of the aquifer = H, hence the discharge can be written as

$$Q = \frac{\pi K(H^2 - h_w^2)}{\ln(R/r_w)}$$

where h_w is the depth of water in the pumping well of radius r_w.

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Example 1

A 30-cm well completely penetrates an unconfined aquifer of saturated depth 40m. After a long period of pumping at a steady rate of 500 lpm, the drawdown in two observation wells 25m and 75m from the pumping well were found to be 3.5m and 2.0m respectively. a) Determine the transmissivity of the aquifer. b) What is the drawdown at the pumping well?

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Ans: a) $T = 3.13 \times 10^{-3}\, m^2/s$ b) $s_w = 11.51\, m$

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Example

Example 2

A fully penetrating well of 30cm diameter in an unconfined aquifer of saturated thickness 50m was found to give the following drawdown-discharge relations under equilibrium condition. If the radius of influence of the well can be assumed to be proportional to the discharge through the well, estimate the flow rate when the drawdown at the well is 6.0m.

Drawdown at the well(m)	Discharge
3.0	600
11.7	1800

Ans:

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Drawdown at the well(m)	Discharge
3.0	600
11.7	1800

Ans: 1512 lpm