GROUNDWATER HYDROLOGY

Groundwater is the water that occurs in a saturated zone of variable thickness and depth below the earth's surface. It is therefore the water beneath the earth's surface from which wells, springs, and groundwater run-off are supplied. Groundwater hydrology may be defined as the science of the occurrence, distribution and movement of water below the surface of the earth. Groundwater referred to without further specification is commonly understood to mean water occupying all the voids within a geological stratum. The main source of groundwater is precipitation. A portion of rain falling on the earth's surface infiltrates into ground travels down and, when checked by impervious layer to travel further down, forms ground water. The ground water reservoir consists of water held in voids within a geologic stratum.

Groundwater may flow into streams, rivers, marshes, lakes and oceans, or it may discharge in the form of springs and flowing wells. Groundwater flows through permeable material, which contains interconnected cracks or spaces that are both numerous enough and large enough to allow water to move freely. In some permeable materials groundwater may travel several meters in a day, in other places, it moves only a few centimeters in a very slowly through relatively impermeable materials such as clay and shale. Thus, the residence time of the groundwater i.e, the length of time water spends in the groundwater portion of the hydrologic cycle, varies enormously.

Groundwater is distinctive from surface water in the following respects: (a) Groundwater exists in voids in the subsurface and its movement is very slow. (b) Pore geometry, soil or rock fracture, surface tension, and flow resistance fundamentally affect groundwater movement, both in saturated and unsaturated conditions. (c) Topographic and geologic structures strictly govern groundwater flows. (d) Soil, stratum, rock mineral and geothermal conditions exert a great influence on the chemical properties of groundwater. Natural topographic and geologic systems control the occurrence of groundwater. Thus groundwater has various types in flow systems based on the topographic and geologic conditions. The water content in the geologic formations varies with depth below ground surface.

Occurrence of groundwater

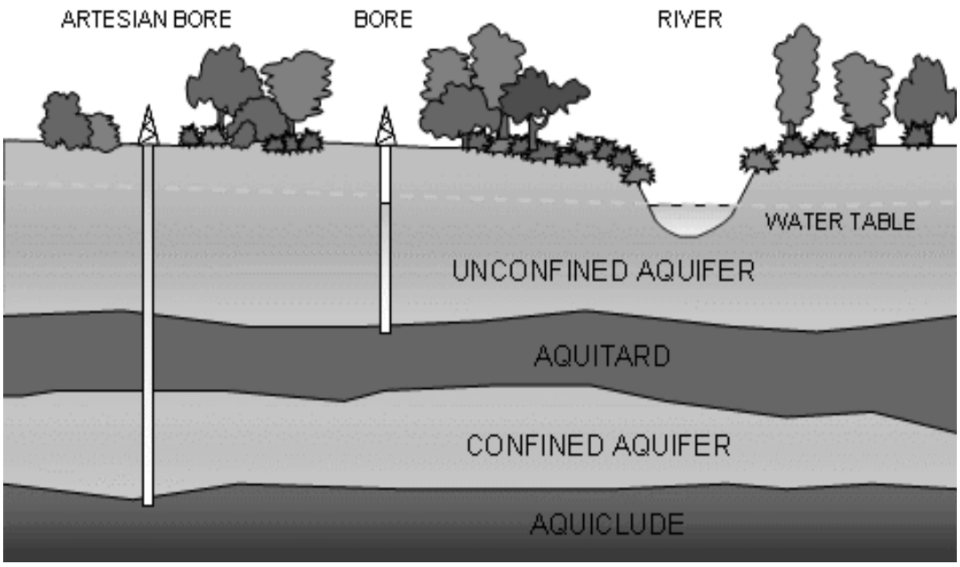
The geologic zones important to groundwater must be identified as well as their structure in terms of water holding and water yielding capabilities. Groundwater occurs in many types of geologic formations known as aquifers.

Aquifer (Groundwater reservoir/water bearing formation)

Formation that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs. Unconsolidated sands and gravels are a typical example. Aquifers are in general areal extensive and may be overlain or underlain by a confining bed which may be defined as relatively impermeable material.

There are various types of confining beds

Aquiclude – A saturated but relatively impermeable material that does not yield appreciable quantity of water to wells. Eg- Clay.



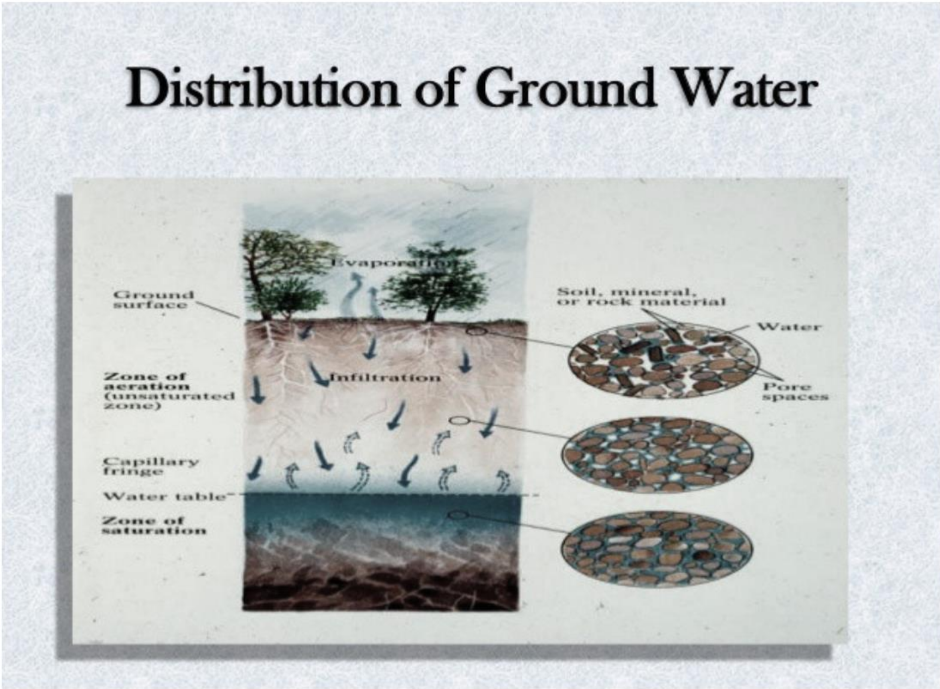
Different types of water bearing stratum

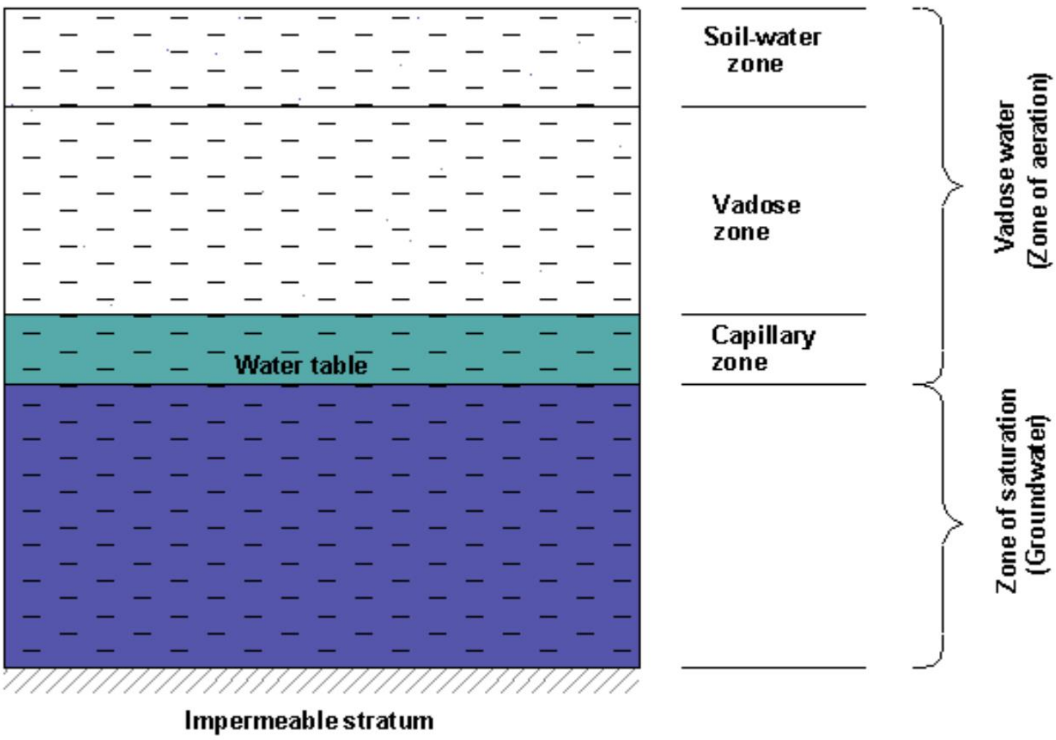
Aquifuge- A relatively impermeable formations neither containing nor transmitting water. Eg- Solid granite

Aquitard- A saturated but poorly permeable stratum that impeds groundwater movement, less yield but it can transmit appreciable amount of water. Eg- Sandy clay

Vertical movement of groundwater

Water, below the surface, at a pressure greater than atmospheric pressure, which thus flows freely into a hole through interconnected void spaces ,is groundwater. .The following Figure illustrates the various zones of water found beneath the surface. Water beneath the surface can essentially be divided into three zones:1) the soil water zone, or vadose zone, 2) an intermediate zone, or capillary fringe and 3) the ground water, or saturated zone. The top two zones, the vadose zone and capillary fringe, can be grouped into the zone of aeration, where during the year air occupies the pore spaces between earth materials.





Sometimes, especially during times of high rainfall, those pore spaces are filled with water. Aeration from the zone of saturation. The elevation of the water table is determined to be where the pore water pressure, Pw, is equal to atmospheric pressure, Pa. The height of the water table will fluctuate with precipitation, increasing in elevation during wet periods and decreasing during dry. In general, the water table has an undulating surface which generally follows the surface topography, but with smaller relief.

The zone of aeration is the region between the earth's surface and the water table. The main components of this region are the soil and rocks. Their pores are at times partly filled with water and air, and aeration occurs when the air and water mix or come into close contact. pose

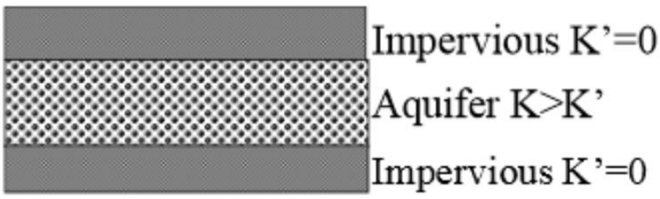
The zone of saturation is the ground immediately below the water table. The pores and fractures in soil and rocks are saturated with water.

1. AQUIFER CLASSIFICATION

Aquifers may be classed as confined, unconfined or leaky which can be taken as a combination of the unconfined and confined.

A confined aquifer is confined above and below by an impervious (may contain water but can't transmit it)layer under pressure greater than the atmospheric. Therefore, in a well penetrating such an aquifer, the water level rises above the bottom of the top confining bed. The water in a confined aquifer is called confined or artesian water. Artesian water flows freely without pumping and the well producing such water is called an artesian or a free flowing well.

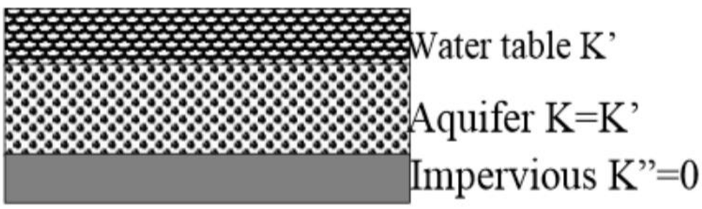
Confined Aquifer (Pressure or piezometric aquifers)



Unconfined Aquifer (Phreatic, Water table)

The water level in well taping an unconfined aquifer and the water table in the aquifer are the same. Therefore, contour maps and profiles of the water table can be prepared from the elevations of water in wells that tap the aquifer to determine the quantities of water available, their distribution and movement.

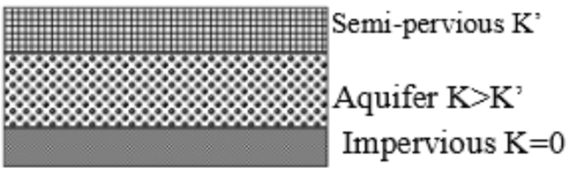
Unconfined Aquifer (Phreatic, Water table)
An unconfined aquifer is one in which a water table (phreatic surface) serves as its upper boundary. A phreatic aquifer is directly recharged from the ground surface above it.



A leaky aquifer

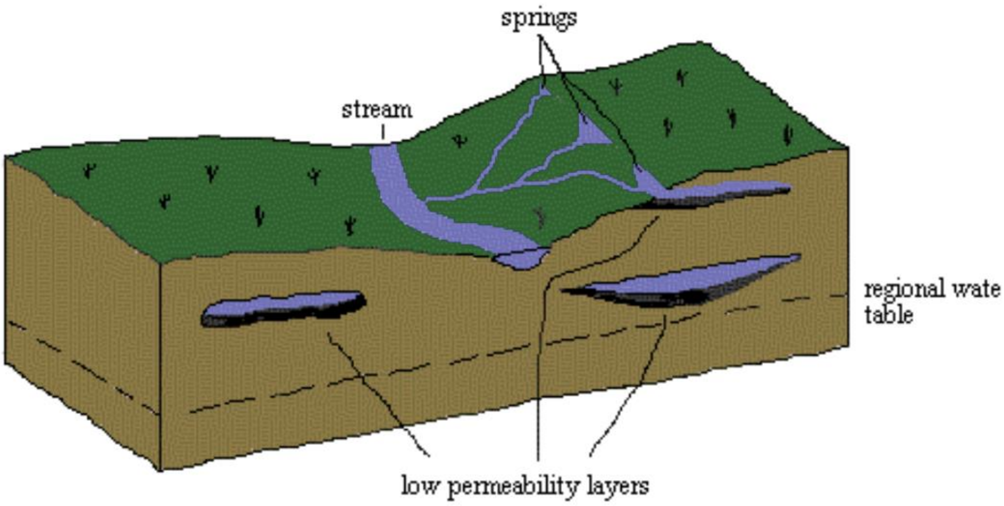
It is an underlain or overlain by semi-pervious strata. Pumping from a well in a leaky aquifer removes water in two ways: by horizontal flow within the aquifer and by vertical leakage or seepage through the semi-confining layer into the aquifer.Aquifers that are completely confined or unconfined occur less frequently than leaky aquifers.

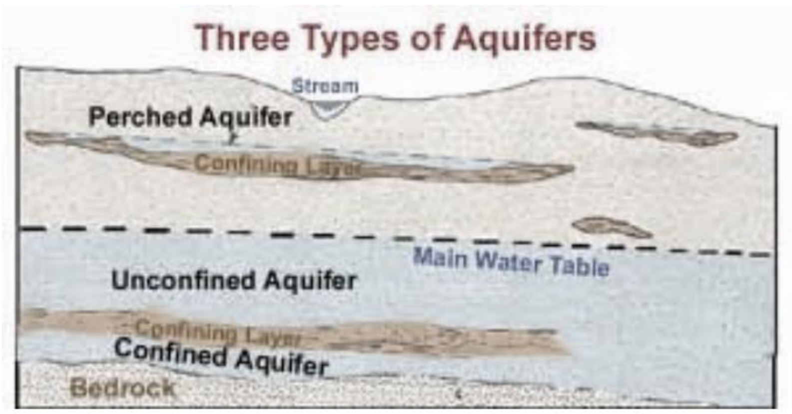
Leaky Aquifer (Semi-confined)



Perched Aquifers

Perched aquifers, are special kinds of phreatic aquifers occurring whenever an impervious (or semi –pervious) layer of limited extent is located between the water table of a phreatic aquifer and the ground surface, thereby making a ground water body, separated from the main groundwater body to be formed. Sometimes, these aquifers exist only during a relatively short part of each year as they drain to the underlying phreatic aquifer. Therefore wells taping such aquifers yield only temporary or small quantities of water.

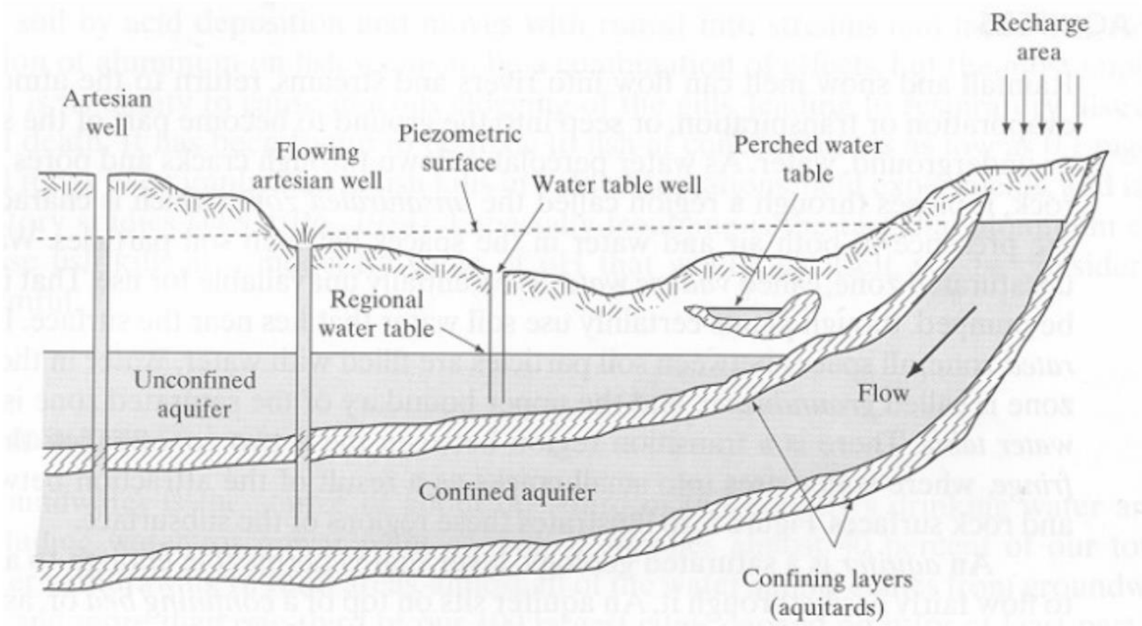




2. Recharge and Discharge

Groundwater recharge represents the portion of rainfall which reaches an aquifer. Groundwater therefore owes its existence directly or indirectly to precipitation. Artificial recharge occurs from excess irrigation seepage from canals and water purposely applied to augment groundwater supplies. Seawater can enter underground along the coasts where the hydraulic gradients slope in an inland direction. The most direct way of quantifying recharge is by examination of borehole hydrograph. This method requires that the specific yield of the aquifer is known. Another approach is to use meteorological data inputs to a recharge simulation model. Natural recharge includes stream bed percolation, deep percolation of rainfall, leakage from ponds, lakes and reservoirs.

Discharge of groundwater occurs when water emerges from underground. Most natural discharge occurs as flows into the surface water bodies e.g. streams, lakes and oceans. Discharge to the ground surface appears as springs. Groundwater discharge also occurs by evaporation from within the soil and by transpiration from vegetation that has access to the water table. However, pumpage from wells constitutes the major artificial discharge of groundwater.



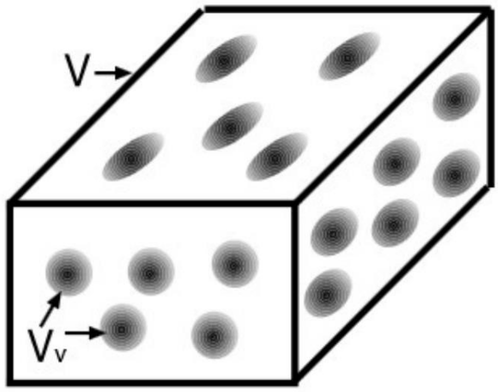
3. Hydrogeological parameters

Porosity

Ground water and soil moisture occur in the cracks, voids, and pore spaces of the Otherwise solid earth materials. Porosity, is the percentage of the total volume that is void of material,

$$n = 100 \times \frac{V_v}{V},$$

where V_v is the volume of void space, and V the total volume. We can write V_v as $V_v = V - V_s$, where the subscript s refers to the volume of the solid phase, $V_s = m_d/\rho_s$, the dry weight divided by the density of the soil/rock.



The term effective porosity refers to the amount of interconnected pore space available for fluid flow and is also expressed as the ratio of the interstices to total volume. Shape, size, packing and degree of cementation affect porosity. Uniformly graded sand has a higher porosity than a less uniform, fine and coarse mixture, because in the latter the fines occupy the voids in the coarse material. In square packing for example, the porosity is as high as 48% while in rhombic packing, it is as low as 26%. Angularity tends to increase porosity while cementation decreases porosity.

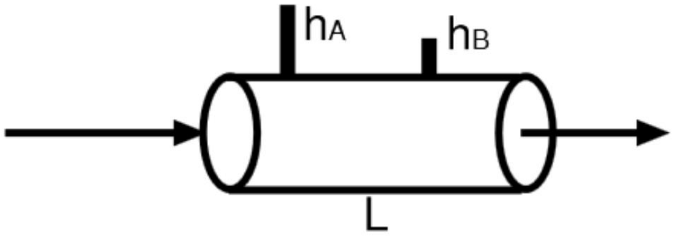
Conductivity and permeability

Some rocks are porous, but the voids are not, or poorly, interconnected. These rockscannot convey water from one void to another. Permeability is the property of an aquifer to transmit water through its pores. The horizontal permeability and vertical permeability may differ.

Co efficient of permeability K is also known as hydraulic conductivity and has the dimensions as those of the velocity of flow. It depends on the fluid property as well as the property of the aquifer soil sample. In the mid-1800s, Henry Darcy, a French engineer, made the first systematic study of the movement of water through a porous medium.
Darcy's law is illustrated in the following Figure, and states that the rate of flow per unit area of an aquifer is proportional to the gradient of potential flow in the direction of flow.

$$q = \frac{Q}{A} = -K\frac{dh}{dl},$$

where q [L T^{-1}] is specific discharge (sometimes called Darcian velocity), A [L^2] is cross-sectional area, Q [L^3 T^{-1}] is discharge, dh/dl is the hydraulic gradient, and *K hydraulic conductivity* [L T^{-1}].



A horizontal pipe filled with sand to demonstrate Darcy's experiment. Water is applied under pressure at point *A* and discharges at point *B*.

Groundwater movement is governed by established hydraulic properties such as porosity, permeability, hydraulic conductivity etc. The ability of an aquifer to receive, store or transmit water or contaminants on the rock properties within the aquifer.

Hydraulic conductivity (unit length per unit time):

The coefficient that describes the ability of a geologic medium to ". . . transmit in unit time a unit volume of ground water at the prevailing viscosity through a cross section of unit area, measured at right angles to the direction of flow, under a hydraulic gradient of unit change in head through unit length of flow." Hydraulic conductivity can be calculated by dividing the transmissivity by the aquifer thickness (Lohman, 1979). Hydraulic Conductivity (K)

K is the specific discharge (v) per unit hydraulic gradient (dh/dl) at a specified temperature and expresses the ease with which fluid is transported through a porous matrix.

$K = v/(dh/dl)$ where $v=Q/A$

It is therefore a coefficient which depends on both matrix and fluid properties. The relevant fluid properties are density ρ and viscosity μ (or in the combined form of kinematic viscosity v). The relevant solid matrix properties are mainly grain- (or pore-) size distribution, shape of grains, arrangement of pores and porosity.

$K = k\rho g/ \mu = kg/ v$

where k (dimensions of L2) – called the permeability, or intrinsic permeability, of the porous matrix – depends solely on properties of the soil matrix. Field measurements of hydraulic conductivity are usually made by carrying out pumping test on wells while laboratory measurements are done using parameters.

Transmissivity (square unit length per unit time): "The rate at which water of the prevailing kinematic viscosity is transmitted [horizontally] through a unit width of the aquifer under a unit hydraulic gradient."
Transmissivity (T) T=Kb where K is the hydraulic conductivity and b, the thickness of the aquifer. T defines the rate at which water of prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient.

Specific yield (unitless): "The ratio of (1) the volume of water which after being saturated, it [rock or soil] will yield by gravity to (2) its [rock or soil] own volume." Specific yield is virtually the same as the storativity for unconfined aquifers.it also known as drainable porosity.

$S_y = V_{wd}/V_t$

S_y - Specific yield

V_{wd}-Volume of water drained

V_t– total rock or material volume

Specific Yield is the volume of water, expressed as a percentage of the total volume of the saturated aquifer that can be drained by gravity. The volume retained (by molecular and surface tension forces) against the force of gravity, expressed as a percentage of the total volume of the saturated aquifer, is called the specific retention.
Porosity = Specific yield + Specific retention.

Specific yield can be determined in the laboratory by simple saturation and drainage, while in the field by pumping a known volume of water out and determining the volume of sediments drained by observing the depth of the water lowered.

Specific yield = (Volume of water pumped out /Volume of sediments drained) x 100

Storage Coefficient or Storativity (unitless): "The volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head.

It defines the volume of water that an aquifer releases from or takes into storage per unit surface area of aquifer per unit change in the component of head normal to that surface. It is a dimensionless quantity involving a volume of water per volume of aquifer.

Specific retention

The ratio of the volume of the water the rock or sediment will retain against the pull of gravity to the total volume of the rock or sediment.

Unconfined aquifer- 0.02 to 0.3
Confined aquifer – 0.00005 to 0.005

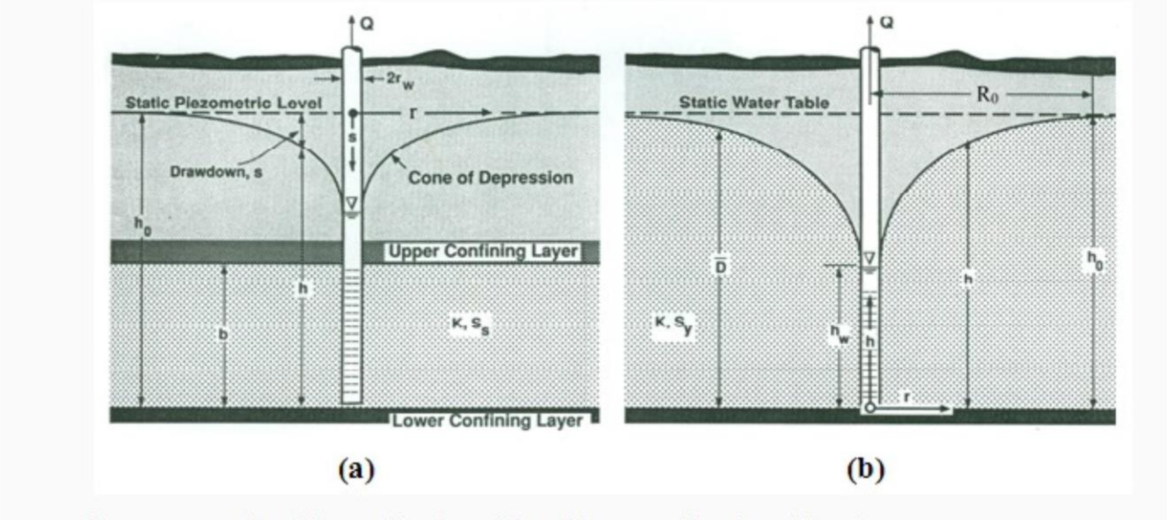
Static water level: is the level of water in the well when no water is being taken out.

Dynamic Water level: is the level when water is being drawn from the well. The cone of depression occurs during pumping when water flows from all directions toward the pump.

4. STEADY STATE FLOW IN TO A WELL

When a well is pumped, water flows into the well from the surrounding aquifer because of difference in hydraulic heads at the well and in the aquifer caused by pumping. Before pumping, water level in the well stands at a height theoretically equal to the static water pressure in the saturated layer around the well. This water level is known as ‘static water level’ or ‘pre-pumping water level’ (Fig. 10.1). When pumping starts, water is removed from the aquifer surrounding the well and the water level in the well ‘piezometric level’ in case of confined aquifers (Fig. 10.1a) and ‘water table’ in case of unconfined aquifers (Fig. 10.1b)] starts lowering. The water level in the well at any instant during pumping is known as ‘pumping water level’.

Steady flow occurs when the water level has ceased to decline as a result of equilibrium between the discharge of the pumped well and the recharge of the aquifer buy outside source. The Dupuit (1863) equation, later modified by Theim, 1906 can be used for analysing steady flow to wells.



Drawdown pattern in: (a) Confined aquifer; (b) Unconfined aquifer.

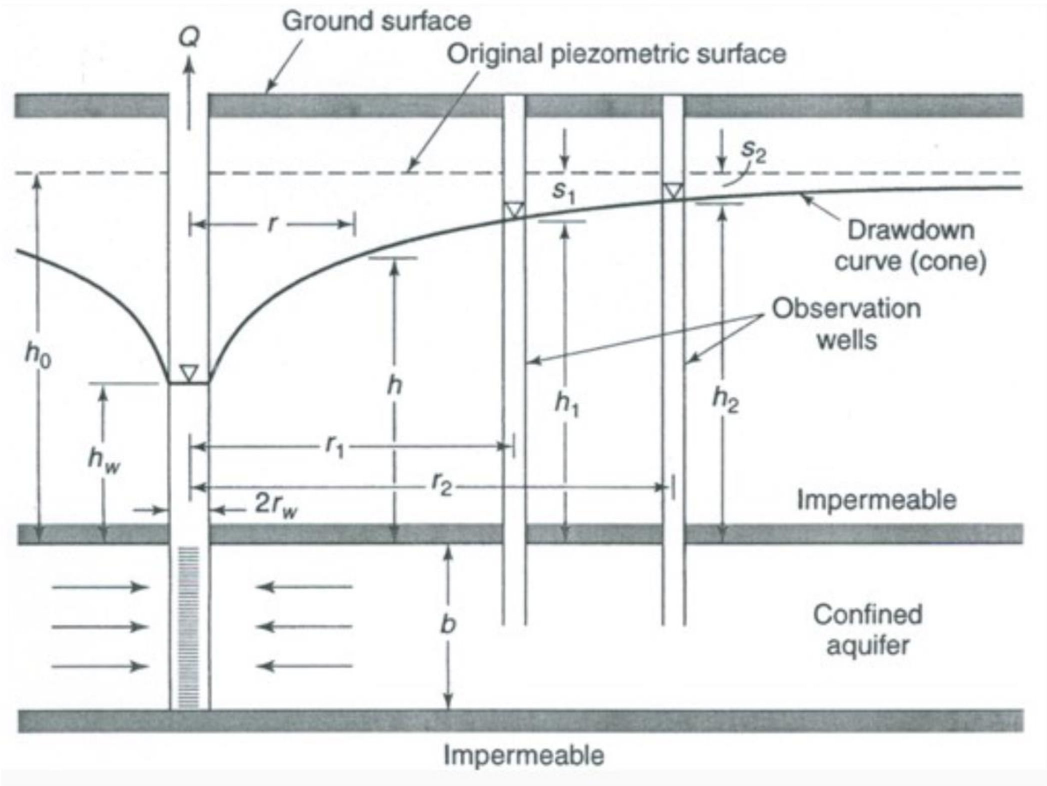
The difference between the static water level and the pumping water level at any instant is called 'drawdown', which is a function of pumping rate, pumping duration and distance from the pumping well. Drawdown is always maximum at the pumping well and it decreases with an increase in the distance from pumping well (Fig. 10.1). The rate of pumping from an aquifer significantly affects the hydraulic gradient in the aquifer. The faster the well is pumped, the steeper the hydraulic gradient will be in the vicinity of the well.

A drawdown curve at a given time shows the variation of drawdown with distance from the pumping well. In three dimensions, the drawdown curve takes the shape of an inverted cone centered on the pumping well, which is known as cone of depression. The outer limit of the cone of depression defines the area of influence of the well. The boundary of the area of influence is called circle of influence and the radius of the circle of influence is called radius of influence. Thus, the radius of influence (R_0) is the distance from a pumping well to the edge of the cone of depression (Fig. 10.1) where drawdown is zero.

As more and more groundwater is pumped through the well, the more water comes from aquifer storage. As a result, the radius of influence increases until when the rate of pumping (discharge) becomes equal to the rate of flow into the well from the area around the well. At this instant of time, a steady flow condition exists in the aquifer and the cone of depression gets stabilized (i.e., it does not change with pumping time). This equilibrium condition changes when the pumping rate is increased or decreased. Note that under steady-state conditions, the entire pumped water is assumed to be coming from external sources beyond the radius of influence. In contrast, under unsteady-state (transient-flow) conditions, either entire pumped water is assumed to be coming from the aquifer storage within the radius of influence or the pumped water is assumed to be coming partly from the aquifer storage within the radius of influence and partly from external sources beyond the radius of influence depending on field conditions.

Confined Aquifers

The following Figure shows a pumping well fully penetrating a confined aquifer and is subjected to pumping. Groundwater level under equilibrium conditions is also depicted. Under equilibrium (steady-state) conditions, the rate of pumping (Q) is equal to the rate that the aquifer transmits water to the well. This problem was first solved by G. Thiem in 1906, which is presented below.



Steady flow to a pumping well in a confined aquifer.

(Source: Todd, 1980)

The yield from the well is

Q = KiA (Darcy's law)

From the Darcy's law, the flow of water through a circular section of the aquifer towards the pumping well is given as:

$$Q = (2\pi rb) \times K \times \frac{dh}{dr}$$

Where, Q = constant rate of pumping from the well, r = radial distance from the circular section to the pumping well, b = thickness of the confined aquifer, K = hydraulic conductivity of the confined aquifer, and dh/dr = hydraulic gradient.

Since transmissivity (T) is the product of aquifer thickness (b) and hydraulic conductivity (K), the above Eqn. can be expressed as:

$$Q = 2\pi r T \times \left(\frac{dh}{dr}\right)$$

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

Let's consider that two observation wells are installed in the aquifer at distances r_1 and r_2 from the pumping well, respectively with hydraulic heads h_1 and h_2.

Rearranging and integrating with appropriate boundary conditions

$$\int_{h_1}^{h_2} dh = \frac{Q}{2\pi \mathrm{T}} \int_{r_1}^{r_2} \frac{dr}{r}$$

$$\Rightarrow \quad h_2 - h_1 = \frac{Q}{2\pi T} \times \ln\left(\frac{r_2}{r_1}\right)$$

$$\therefore \quad Q = 2\pi T \times \frac{h_2 - h_1}{\ln\left(\frac{r_2}{r_1}\right)}$$

In practice, instead of the hydraulic head (h), drawdown (s) is measured, and hence after expressing h_1 and h_2 as draw downs, Eqn. can be written as:

$$Q = 2\pi \mathrm{T} \times \frac{(s_1 - s_2)}{\ln\left(\frac{r_2}{r_1}\right)}$$

Where, s_1 and s_2 are steady draw downs in the two observation wells located respectively at r_1 and r_2 distances from the pumping well.

Furthermore, if we consider that the first observation well is located at a distance r_w (radius of the pumping well) where hydraulic head is h_w and instead of the second observation well at r_2, we consider that $r_2 = R_0$ (radius of influence) where drawdown (s_2) is zero and hence hydraulic head is h_0 (static or pre-pumping groundwater level), the Thiem equation can be expressed as follows:

$$Q = 2\pi T \times \frac{h_0 - h_w}{\ln\left(\frac{R_0}{r_w}\right)}$$

$$Q = 2\pi T \times \frac{s_w}{\ln\left(\frac{R_0}{r_w}\right)}$$

The above is referred to as the equilibrium or Theim's equation. r_1, r_2 are respective distances of piezometers (observation wells) to the pumped well h_2, h_1 are their respective water levels. The Theim's equation enables the K or T of an aquifer to be determined from a pumped well being monitored from at least two observation wells at different distances from the pumped well.

If the drawdown in the observation wells are s_1 and s_2, then h_2-h_1 = s_1-s_2.

The assumptions in **Theim's equation** are:

i. the aquifer is confined.

ii. the aquifer has an infinite areal extent.

iii. the aquifer is homogeneous, isotropic and of uniform thickness over the area influenced by the test.

iv. prior to pumping, the piezometric surface is horizontal over the area that will be influenced by the test.

v. the aquifer is pumped at a constant discharge rate.

vi. the well penetrates the entire saturated thickness of the aquifer.

vii. the flow to the well is at steady state.

Unconfined Aquifers

The analysis of flow in unconfined aquifers is more complicated than that in confined aquifers. Thiem also derived an equation for steady radial flow in unconfined aquifers which is discussed in this section. Besides the basic assumptions and the assumptions of radial symmetry and steady-state condition mentioned above, the following additional assumptions are made in this case:

(1) The aquifer is unconfined and underlain by a horizontal confining layer.

(2) The well is pumped at a constant rate.

(3) The Dupuit-Forchheimer assumptions are valid.

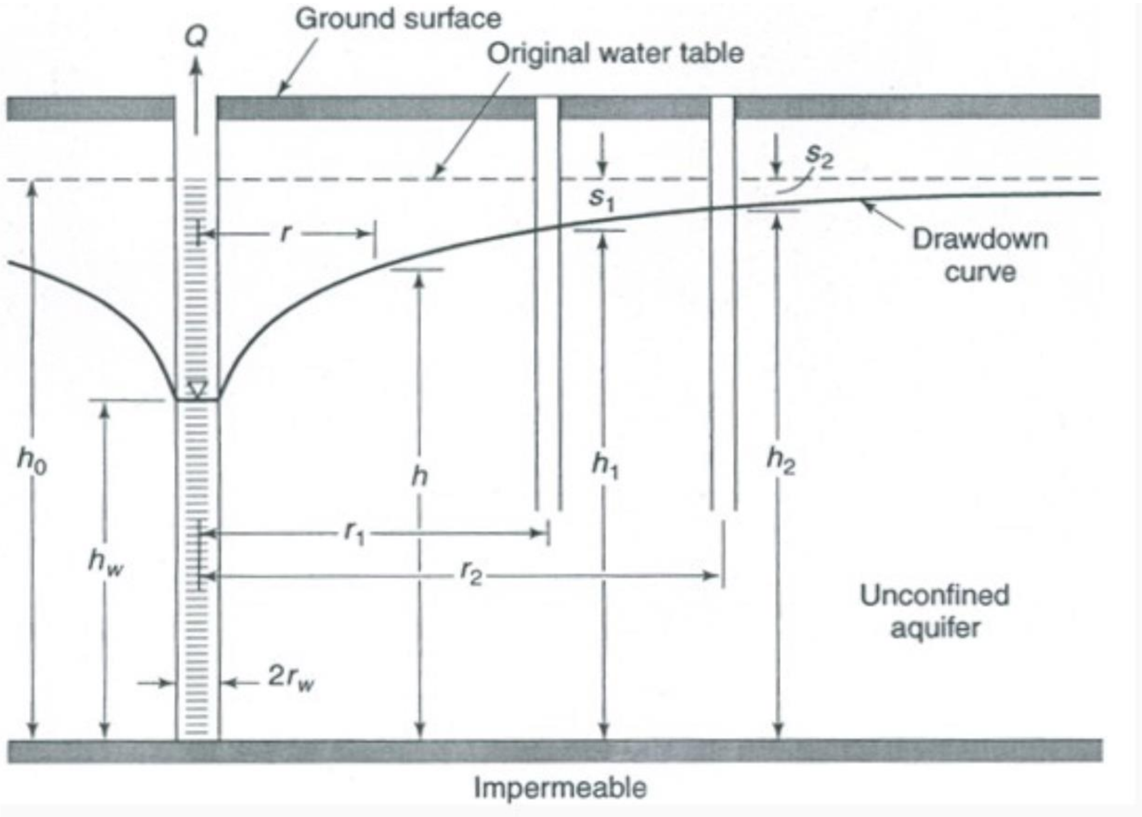
Using the Darcy's law, the radial flow in the unconfined aquifer (Fig. 10.3) can be described as:

Q = KiA (**Darcy's law**)

Where, Q = Constant rate of pumping, r = radial distance from the circular section to the pumping well, h = saturated thickness of the unconfined aquifer, K = hydraulic conductivity of the unconfined aquifer, and dh/dr = hydraulic gradient.

$$Q = (2\pi rh) \times K \times \frac{dh}{dr}$$

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



Steady flow to a pumping well in an unconfined aquifer

Let's consider that two observation wells are located in the unconfined aquifer at distances r_1 and r_2, with hydraulic heads h_1 and h_2, respectively (Fig. 10.3). Now, Eqn. (10.10) can be integrated with these boundary conditions as:

$$\int_{h_1}^{h_2} h\,dh = \frac{Q}{2\pi K}\int_{r_1}^{r_2}\frac{dr}{r}$$

$$\Rightarrow \quad h_2^2 - h_1^2 = \frac{Q}{\pi K}\times \ln\left(\frac{r_2}{r_1}\right)$$

$$\therefore \quad Q = \pi K \times \frac{h_2^2 - h_1^2}{\ln\left(\frac{r_2}{r_1}\right)}$$

Where the drawdown is appreciable, the heads h1 and h2 can be replaced by (h0 - s1) and (h0-s2) respectively.h0 is the original water table. In unconfined aquifers, the Dupuit-Theim also assume that the slope of the phreatic surface is very small, therefore flow is considered to be purely horizontal and also uniformly distributed with depth.

Dupuit's equilibrium formula-Unconfined aquifer

$$Q = K.I.A.$$

$$Q = K\cdot\frac{dh}{dr}\cdot 2\pi r h$$

where K = Coefficient of permeability

$$\frac{dr}{r} = \frac{2\pi K}{Q}\cdot h\cdot dh.$$

Integrating between the limits r_w and R, we get

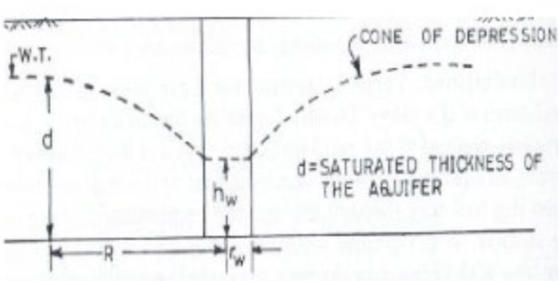
$$\int_{r_w}^{R}\frac{dr}{r} = \frac{2\pi K}{Q}\int_{h_w}^{d} h.dh$$

$$\left|\log_e r\right|_{r_w}^{R} = \frac{2\pi K}{Q}\left|\frac{h^2}{2}\right|_{h_w}^{d}$$

$$\log_e\frac{R}{r_w} = \frac{\pi K}{Q}\cdot[d^2 - h_w^{\,2}]$$

$$2.3\log_{10}\frac{R}{r_w} = \frac{\pi K}{Q}[d^2 - h_w^{\,2}]$$

$$Q = \frac{\pi K\,(d^2 - h_w^{\,2})}{2.3\log_{10}\frac{R}{r_w}}$$



Confined aquifer

$$Q = KIA$$

or

$$Q = K \cdot \frac{dh}{dr} 2\pi \cdot rH$$

or

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

Integrating between r_w and R, we get

$$\int_{r_w}^{R} \frac{dr}{r} = \frac{2\pi K}{Q} \cdot H \int_{h_w}^{D} dh$$

or

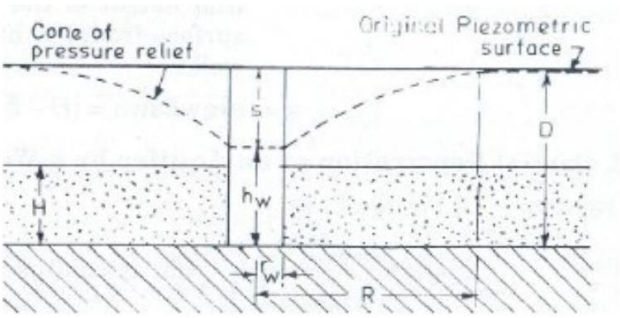
$$\left| \log_e r \right|_{r=r_w}^{r=R} = \frac{2\pi K}{Q} \cdot H \left| h \right|_{h=h_w}^{h=D}$$

or

$$2.3 \log_{10} \frac{R}{r_w} = \frac{2\pi KH}{Q} [(D - h_w)]$$

$$K = \frac{2.3\, Q \log_{10} \frac{R}{r_w}}{2\pi.H.\,(D - h_w)}$$

$$\boxed{Q = \frac{2\pi KH\,(D - h_w)}{2.3 \log_{10} \frac{R}{r_w}}}$$



Numerical Problems

1. **A 30cm well fully penetrates a confined aquifer 30 m deep.After a long period of pumping at a rate of 1,200 lpm,the drawdowns in the wells at 20 and 45 m from the pumping well are found to be 2.2 and 1.8 m respectively. Determine the transmissibility of the aquifer.what is the drawdown in the pumped well?**

Solution

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

Q= 1.2 m^3/min; s_1=2.2m; s_2= 1.8m; r_1= 20m; r_2= 45m

1.2= 2x3.14xTx(2.2-1.8)/[2.303$\log_{10}$(45/20)]
T= 559$m^{2/}$day

2. **A 30 cm well penetrates 50 m below the static water table. After a long period of pumping at a rate of 1800 lpm, the drawdown in the wells at 15 and 45 m from the pumped well were 1.7 and 0.8 m, respectively. Determine the transmissibility of the aquifer. what is the drawdown in the pumped well?**

Solution

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

S_1= 1.7m;H1=48.3m; s_2= 0.8m; h_2= 49.2m; r_1= 15m; r_2=45m;Q=1.8 m^3/min or Q=1.8x60x24 m^3/day; H=50m

By applying the above eqn
K= 10.07 m/day; T= KxH
T= 10.07x 50=**503m^2/day**

Drawdown in the pumped well

To find out the drawdown in the pumped well, the eqn can be rewritten as

Q= 3.14xKx $(h_2^2 - h_w^2)/\ln(r_2/r_w)$ (applying the eqn between well 2 and pumping well)

Find h_w

h_w = 44.19m

To find out drawdown in the pumped well (s_w)

s_w= $(h_2 - h_w) + s_2$ = **5.81m**

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