

GROUND WATER

Types of aquifers and formations

Ground water is widely distributed under the ground and is a replenishable resource unlike other resources of the earth. The problems in Ground Water Investigation are the zones of occurrence and recharge. A water bearing geologic formation or stratum capable of transmitting water through its pores at a rate sufficient for economic extraction by wells is called **'aquifer'**.

Formations that serve as good aquifers are:

- Unconsolidated gravels, sands, alluvium
- Lake sediments, glacial deposits
- Sand stones
- Limestone's with cavities (caverns) formed by the action of acid waters
- Granites and marble with fissures and cracks, weathered gneisses and schist

A geologic formation, which can absorb water but cannot transmit significant amounts is called an **'aquiclude'**. Examples are clays, shale's, etc.

A geologic formation with no interconnected pores and hence can neither absorb nor transmit water is called an **'aquifuge'**. Examples are basalts, granites, etc.

A geologic formation of rather impervious nature, which transmits water at a slow rate compared to an aquifer (insufficient for pumping from wells) is called an **'aquitard'**. Examples are clay lenses interbedded with sand.

CONFINED AND UNCONFINED AQUIFERS

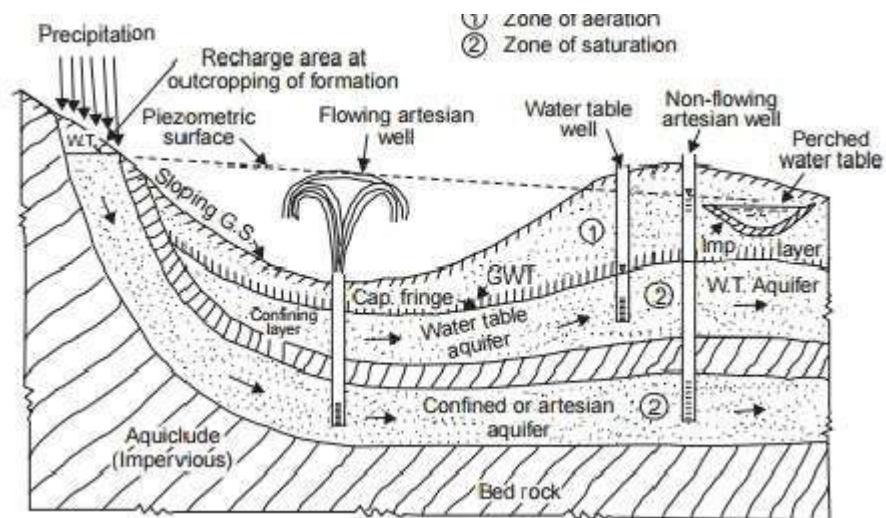


Figure 1 Types of aquifers and location of wells

If there is homogeneous porous formation extending from the ground surface up to an impervious bed underneath (Fig. 5.1), rainwater percolating down in the soil saturates the formation and builds up the ground water table (GWT). This aquifer under water table conditions is called an

unconfined aquifer (water-table aquifer) and well drilled into this aquifer is called a water table well.

On the other hand, if a porous formation underneath is sandwiched between two impervious strata (aquicludes) and is recharged by a natural source (by rain water when the formation outcrops at the ground surface—recharge area, or outcrops into a river-bed or bank) at a higher elevation so that the water is under pressure in the aquifer (like pipe flow), i.e., artesian condition. Such an aquifer is called an artesian aquifer or confined aquifer. If a well is drilled into an artesian aquifer, the water level rises in the well to its initial level at the recharge source called the piezometric surface. If the piezometric surface is above the ground level at the location of the well, the well is called ‘flowing artesian well’ since the water flows out of the well like a spring, and if the piezometric surface is below the ground level at the well location, the well is called a non-flowing artesian well. In practice, a well can be drilled through 2-3 artesian aquifers (if multiple artesian aquifers exist at different depths below ground level). Sometimes a small band of impervious strata lying above the main ground water table (GWT) holds part of the water percolating from above. Such small water bodies of local nature can be exhausted quickly and are deceptive. The water level in them is called ‘**perched water table**’

ACQUIFIER PARAMETERS

Specific yield : While porosity (n) is a measure of the water bearing capacity of the formation, all this water can not be drained by gravity or by pumping from wells as a portion of water is held in the void spaces by molecular and surface tension forces. The volume of water expressed as a percentage of the total volume of the saturated aquifer, that will drain by gravity when the water table (Ground Water Table (GWT) drops due to pumping or drainage, is called the ‘specific yield (S_y)’ and that percentage volume of water, which will not drain by gravity is called ‘specific retention (S_r)’ and corresponds to ‘field capacity’ i.e., water holding capacity of soil (for use by plants and is an important factor for irrigation of crops). Thus, porosity = specific yield + specific retention $n = S_y + S_r$. Specific yield depends upon grain size, shape and distribution of pores and compaction of the formation. The values of specific yields for alluvial aquifers are in the range of 10–20% and for uniform sands about 30%

Storage coefficient: The volume of water given out by a unit prism of aquifer (i.e., a column of aquifer standing on a unit horizontal area) when the piezometric surface (confined aquifers) or the water table (unconfined aquifers) drops by unit depth is called the storage coefficient of the aquifer (S) and is dimensionless (fraction). It is the same as the volume of water taken into storage by a unit prism of the aquifer when the piezometric surface or water table rises by unit depth. In the case of water table (unconfined) aquifer, the storage coefficient is the same of specific yield (S_y). Since the water is under pressure in an artesian aquifer, the storage coefficient of an artesian aquifer is attributable to the compressibility of the aquifer skeleton and expansibility of the pore water (as it comes out of the aquifer to atmospheric pressure when the well is pumped) and is given by the relationship

$$S = \gamma_w n b \left(\frac{1}{K_w} + \frac{1}{n E_s} \right)$$

S = storage coefficient (decimal)

γ_w = specific weight of water

n = porosity of soil (decimal)

b = thickness of the confined aquifer

K_w = bulk modulus of elasticity of water

E_s = modulus of compressibility (elasticity) of the soil grains of the aquifer.

The storage coefficient of an artesian aquifer ranges from 0.00005 to 0.005, while for a water table aquifer $S = S_y = 0.05-0.30$.

Porosity (n): Those portions of soil, not occupied by solids; Ratio of volume of pores or interstices to total volume.

Permeability – an expression of movement of water in any direction

Vertical Distribution of ground water:

Water in the subsurface may be divided into two major zones: i) water stored in the unsaturated zone also known as vadose zone or zone of aeration and ii) water stored in the saturated zone. Soil pore spaces in the vadose zone, lying immediately below the surface. Here the small pore spaces between soil particles are filled with a mixture of water and air resulting in an area which is less than saturated zone. This zone may be divided with respect to occurrence and circulation of water into the uppermost zone of soil water, the intermediate zone and the capillary fringe, immediately above the water table. Water in this zone is called capillary water. This water moves upward from the water table by capillary action. Capillary water moves slowly in any direction. Water cannot be withdrawn from this zone for residential or commercial water supply purpose because the capillary forces hold it too tightly.

The roots of trees, plants and crops, however, can tap into this water. The capillary fringe moves upwards and downwards together with the water table due to seasonal pattern. Fig shows the distribution of water in the subsurface regions. Groundwater is water below the water table, filling entirely all rock interstices (void spaces) in the saturated zone. The water located in this zone can be withdrawn for various uses. The variation in the flow of groundwater depends on the type of rocks or other permeable material, the size of the pore spaces in the soil or rock, connectivity of pore spaces, and the configuration of the underground strata. Water Table: The upper surface of the zone of saturation is known as water table. At the water table, the water in the pores of the aquifer is at atmospheric pressure.

The hydraulic pressure at any level within a water table aquifer is equal to the depth from the water table point and is referred to as the hydraulic head. When a well is dug in a water table aquifer, the static water level in the well stands at the same elevation as the water table. The groundwater table, sometimes called the free or phreatic surface, is not a stationary surface. This water table moves up and down due to various reason. It may rises when more water is added to the saturated zone by vertical percolation, and drops down during drought periods when the stored water flows out towards springs, streams, well and other points of groundwater discharge.

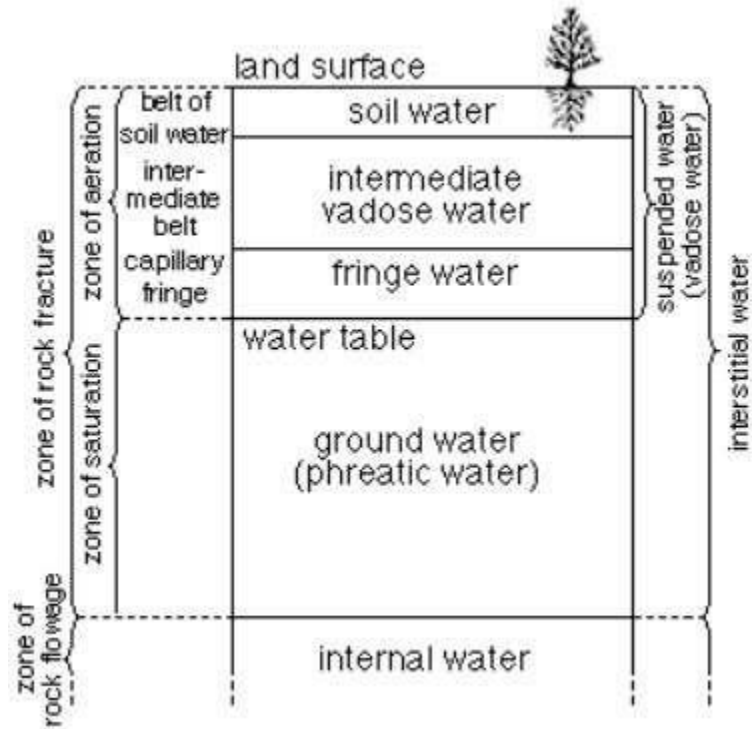


Fig: Subsurface Distribution of water

DARCY'S LAW

Flow of ground water except through coarse gravels and rockfills is laminar and the velocity of flow is given by Darcy's law (1856), which states that 'the velocity of flow in a porous medium is proportional to the hydraulic gradient', Fig. 5.2

$$V = Ki, \dots(1)$$

$$i = \Delta h / L \dots(2)$$

$$Q = AV = AKi, \dots(3)$$

$$A = Wb, T = Kb$$

$$Q = WbKi$$

$$Q = T iw \dots(4)$$

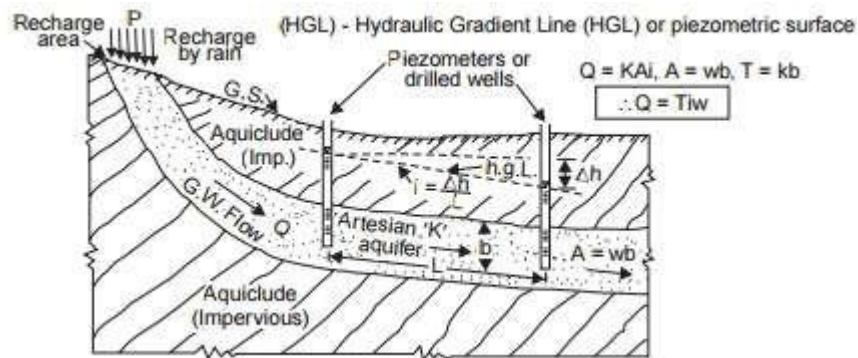


Figure 2 : Flow of ground water

where V = velocity of flow through the aquifer

K = coefficient of permeability of aquifer soil

$i = \text{hydraulic gradient} = \Delta h / L$,

$\Delta h = \text{head lost in a length of flow path } L$

$A = \text{cross-sectional area of the aquifer} (= wb)$ $w = \text{width of aquifer}$ $b = \text{thickness of aquifer}$

$T = \text{coefficient of transmissibility of the aquifer}$

$Q = \text{volume rate of flow of ground water (discharge or yield)}$

Darcy's law is valid for laminar flow, i.e., the Reynolds number (Re) varies from 1 to 10, though most commonly it is less than 1

TRANSMISSIBILITY

It can be seen from Eq. (4) that $T = Q$, when $i = 1$ and $w = 1$; i.e., the transmissibility is the flow capacity of an aquifer per unit width under unit hydraulic gradient and is equal to the product of permeability times the saturated thickness of the aquifer. In a confined aquifer, $T = Kb$ and is independent of the piezometric surface. In a water table aquifer, $T = KH$, where H is the saturated thickness. As the water table drops, H decreases and the transmissibility is reduced. Thus, the transmissibility of an unconfined aquifer depends upon the depth of GWT.

WELL HYDRAULICS

Steady radial flow into a well (Dupuit 1863, Thiem 1906)

Assuming that the well is pumped at a constant rate Q for a long time and the water levels in the observation wells have stabilised, i.e., equilibrium conditions have been reached, Fig. 5.3 (a).

From Darcy's law,

$$Q = K i A$$

$$Q = K (2\pi xy)$$

$$Q = 2\pi K$$

$$Q = \frac{\pi K (h_2^2 - h_1^2)}{2.303 \log_{10} \left(\frac{r_2}{r_1} \right)} \dots\dots 6$$

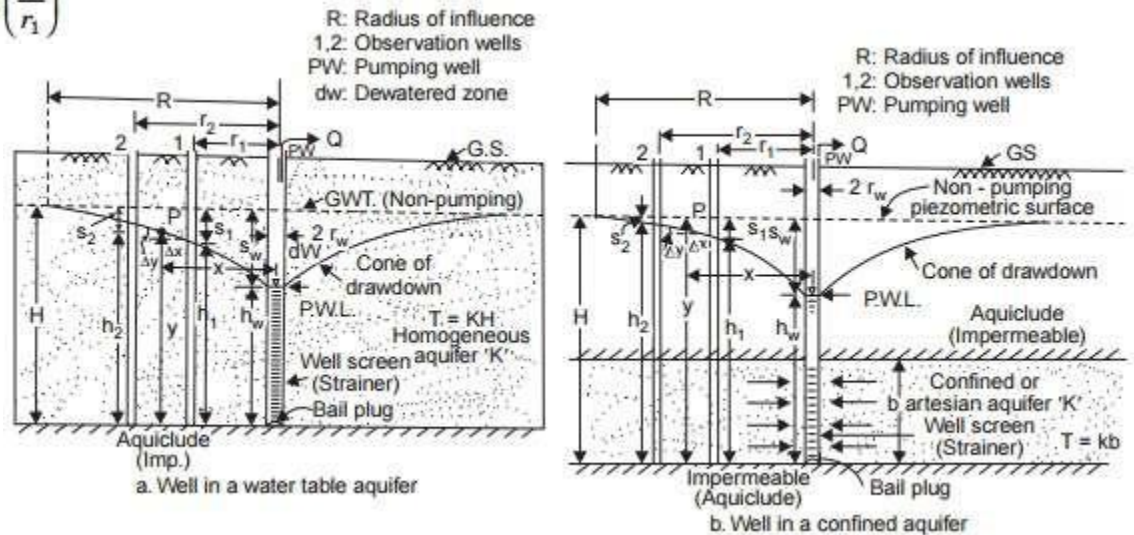


Figure 3: Steady radial flow into a well

Applying the Eq. (6) between the face of the well ($r = r_w$, $h = h_w$) and the point of zero drawdown ($r = R$, $h = H$)

$$Q = \frac{\pi K(H^2 - h_w^2)}{2.303 \log_{10} \left(\frac{R}{r_w} \right)} \quad \dots\dots\dots 7$$

If the drawdown in the pumped well ($S_w = H - h_w$) is small
 $H^2 - h_w^2 = (H + h_w)(H - h_w)$, $H + h_w \approx 2H$

$$= 2H(H - h_w)$$

$$Q = \frac{2\pi KH(H - h_w)}{2.303 \log_{10} \left(\frac{R}{r_w} \right)}, \quad \dots\dots\dots 8$$

(b) Artesian conditions (confined aquifer)

If the well is pumped at constant pumping rate Q for a long time and the equilibrium conditions have reached, Fig. 5.3 (b).

From Darcy's law, $Q = K i A$

$$Q = K (2\pi x b)$$

$$Q = 2\pi K$$

$$Q = \frac{2\pi (kb)(h_2 - h_1)}{2.303 \log_{10} \left(\frac{r_2}{r_1} \right)} \quad \dots\dots\dots 9$$

Applying Eq. (7.9) between the face of the well ($r = r_w$, $h = h_w$) and the point of zero drawdown ($r = R$, $h = H$), simplifying and putting $T = Kb$,

$$Q = \frac{2.72 T (H - h_w)}{\log_{10} \left(\frac{R}{r_w} \right)}$$

Dupuit's Equations Assumptions:

The following assumptions are made in the derivation of the Dupuit Thiem equations:

- (i) Stabilized drawdown—i.e., the pumping has been continued for a sufficiently long time at a constant rate, so that the equilibrium stage of steady flow conditions have been reached.
- (ii) The aquifer is homogeneous, isotropic, of infinite areal extent and of constant thickness, i.e., constant permeability.
- (iii) Complete penetration of the well (with complete screening of the aquifer thickness) with 100% well efficiency.
- (iv) Flow lines are radial and horizontal and the flow is laminar, i.e., Darcy's law is applicable.
- (v) The well is infinitely small with negligible storage and all the pumped water comes from the aquifer.

SPECIFIC CAPACITY

The specific capacity Q/S_w of a well is the discharge per unit drawdown in the well and is usually expressed as lpm/m. The specific capacity is a measure of the effectiveness of the well; it decreases with the increase in the pumping rate (Q) and prolonged pumping (time, t). In Eq. (8) by putting $r_w = 15$ cm, $R = 300$ m, $KH = T$ $H - h_w = S_w$, the specific capacity

$$= T/1.2$$

HYDRAULICS OF OPEN WELLS

This equation does not apply for shallow dug open wells since there is no instantaneous release of water from the aquifer, most of the water being pumped only from storage inside the well (Fig. 5.4).

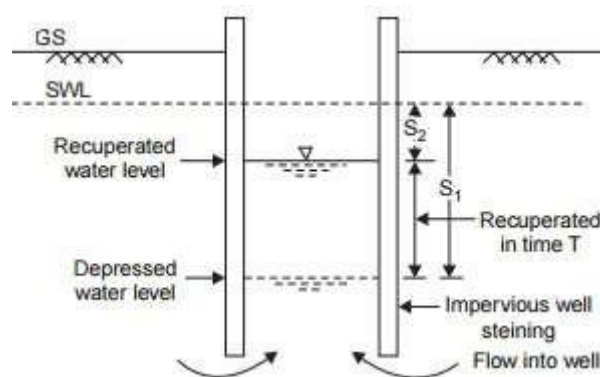


Figure 4: Recuperation test in open wells

In alluvial soil, if the water is pumped at a high rate the depression head (static water level–water level inside the well during pumping) will increase, which may cause excess gradients resulting in loosening of sand particles (quick sand phenomenon). This limiting head is called ‘critical depression head’. The ‘safe working depression head’ is usually one-third of the critical head and the yield under this head is called the maximum safe yield of the well.

Yield Tests

The following tests may be performed to get an idea of the probable yield of the well: (a) Pumping test (b) Recuperation test

(b) Recuperation Test. In the recuperation test, the water level in the well is depressed by an amount less than the safe working head for the subsoil. The pumping is stopped and the water level is allowed to rise or recuperate. The depth of recuperation in a known time is noted from which the yield of the well may be calculated as follows (Fig. 5.4).

Let the water level inside the well rise from s_1 to s_2 (measured below static water level, swl) in time T . If s is the head at any time t , from Darcy’s law

$$Q = Kai$$

if a head s is lost in a length L of seepage path

$$Q = KA$$

$$Q = CA_s$$

where the constant $C = K/L$ and has dimensions of T^{-1} .

If in a time dt , the water level rises by an amount ds $Q dt = - A ds$

the -ve sign indicates that the head decreases as the time increases. Putting $Q =$

$$CA \frac{ds}{dt} = - A ds$$

C

T

dt

=

C

=

Assuming the flow is entirely from the bottom (impervious steining of masonry), the yield of the well

$$Q = CAH$$

From this equation, if $Q = C$ when $A = 1$, $H = 1$, i.e., the specific yield of the soil is the discharge per unit area under a unit depression head and has dimension of T^{-1} (1/time) and the usual values are

$$C = 0.25 \text{ hr}^{-1} \text{ for}$$

$$\text{clayey soil } C = 0.50$$

$$\text{hr}^{-1} \text{ for fine sand}$$

$$C = 1.00 \text{ hr}^{-1} \text{ for coarse sand}$$