CONCEPTS OF GROUNDWATER HYDROLOGY

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General

Groundwater is one of the important components of hydrologic cycle and can be defined as water below the water table where all of the pore spaces are filled with water. Ground water is generally a reliable, high quality resource, but as with any natural resource care must be taken to protect it against pollution threats. In India, groundwater serves about 80% of rural population, 50% of urban population and about 60% of agricultural area. There are more than 20 million groundwater extraction structures in place which are being used to meet requirement for domestic, industrial, and agricultural activities. The demand of groundwater is continuously in the rise because of its some fascinating features, such as; slow moving, large storage volume and long retention time, can be drawn on demand, less risk free than surface water sources, etc. The dependence on groundwater as a reliable source to meet the requirements for irrigation, drinking and industrial uses in India has been risen rapidly during the last few decades. This has resulted in depletion of groundwater table in many areas causing concerns for the long-term sustainability of groundwater based supplies. This short lecture note deals with basic concept of groundwater hydrology.

Subsurface Water

The subsurface water has been divided in two zones: unsaturated or vadose zone and saturated zone (Figure 1). The unsaturated zone is further subdivided into three zones, *i.e.,* soil water zone, intermediate zone and capillary zone. The soil water zone is adjacent to the ground surface. The intermediate zone is between the lower edge of the soil water zone and the upper edge of the capillary zone. The capillary zone extends from the bottom edge of the intermediate zone to the upper edge of the saturated zone. The thickness of the capillary zone depends on the properties of the soil and also on the homogeneity of the soil. The depth of capillary zone is varying from few centimetres to few meters. Saturated zone starts from the bottom edge of the capillary zone. In this zone, all the pores of the soil matrix are filled with water. The water contained in this zone is called as Groundwater. The top surface of the zone of saturation or groundwater is known as phreatic surface or water table.

Aquifer Systems

Aquifers are bodies of consolidated or unconsolidated deposits that store and transmit water in useable quantities. Aquifers are principally found in one of the four categories of geological formations

(Jacob, 2004):

- Unconsolidated (sand and gravels)
- Permeable sedimentary rocks (sandstones)
- Carbonate rocks (limestone and dolomite)
- Igneous rocks (heavily fractured volcanic and crystalline rocks).

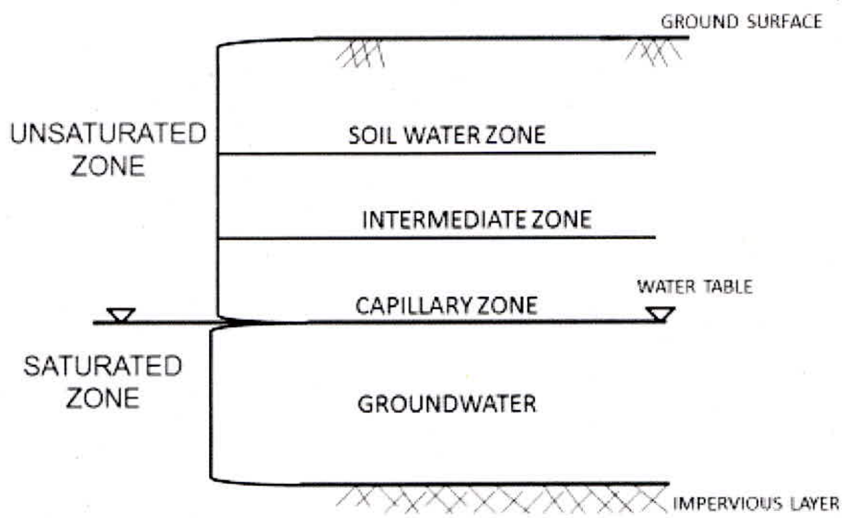

Figure 1: Classification of subsurface water

Aquitards are formations which are capable of storing water but have relatively low permeability. Generally aquitards consist of fine grained soils like silt or sandy clay. Aquicludes are similar in nature to aquitards, but with even lower permeability such that flow through aquicludes can generally be assumed to be negligible. Aquifuges, generally made up of solid rock, do not contain any interconnected pore space, and thus are impermeable to flow.

Aquifers may be identified as confined, unconfined, or semi-confined depending on the nature of their upper surface or boundaries. The three basic aquifer types are shown schematically in Figure 2. Aquifers that are bounded by an aquifuge or aquiclude are referred to as confined aquifers. Typically the upper confining layer contains clay or other fine grained deposits. By definition, in a confined aquifer, the elevation of the hydraulic head exceeds the top of the upper confining layer. In an artesian aquifer, the hydraulic head exceed the elevation of the ground surface. The water level elevation recorded in a piezometer penetrating a confined aquifer is known as the "piezometric" or "potentiometric level". In an artesian aquifer, where the piezometric surface exceeds the land surface "flowing wells" may result.

Unconfined or phreatic aquifers have an upper water level surface (also known as the water table or phreatic surface) which is at atmospheric pressure. Immediately above the water table of an unconfined aquifer the porous media remains saturated due to capillary action, but the water pressure is less than the ambient air pressure. Semi-confined or leaky aquifers result when the upper confining layer in a semi-confined aquifer is sufficiently permeable to allow flow of water between it and overlying or underlying aquifers.

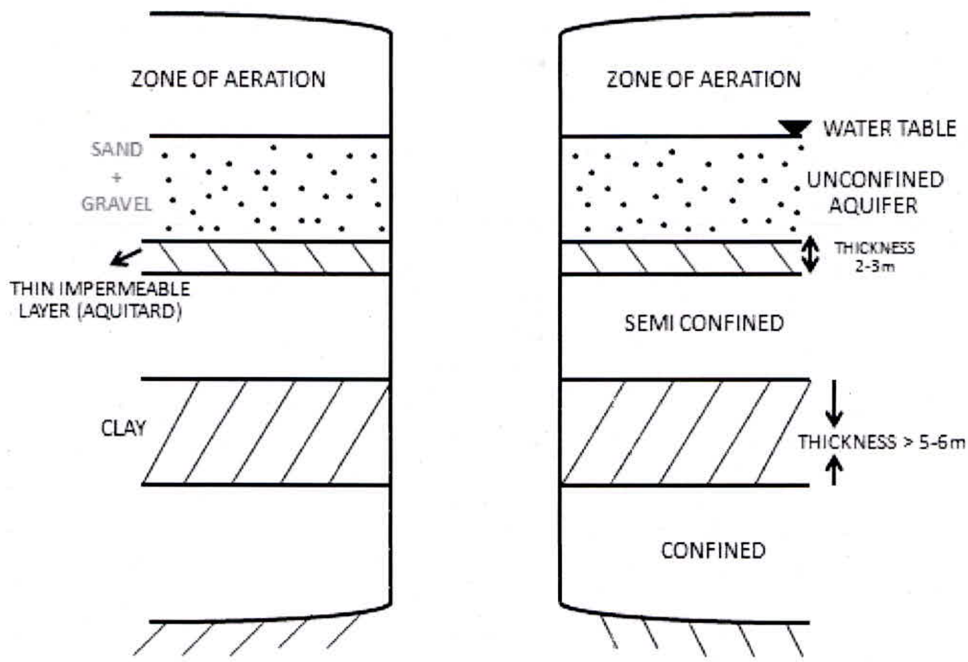

Figure 2: Schematic diagram showing different aquifer systems

Homogeneous and Isotropic Medium

A porous medium is called homogeneous when aquifer parameters are constant throughout the medium, ***i.e.*** the properties of the medium are independent of space. The medium will be called non-homogeneous when aquifer properties are varying with space.

A porous medium will be called isotropic when medium parameters are constant in all the directions, ***i.e.*** the parameters are independent of direction ($Kx = Ky = Kz$). The medium will be called an anisotropic when the parameters are different in different directions ($Kx \neq Ky \neq Kz$).

Porous Media

A major component of precipitation that falls on the earth surface eventually enters into the ground by the process of infiltration. The infiltrated water is stored in the pores of the underground soil strata. When all the pores of a soil matrix are filled with water, we call that the soil is in the state of saturation. We used the term *porosity* to quantify the amount of voids space available in a soil matrix. Porosity is defined as the ratio of volume of voids to the total volume of the soil matrix. The porosity is expressed as,

$$n = Vv/Vt$$

Where, Vv is the volume of voids

Vt is total volume of material

Porosity is a dimensionless parameter and ranges between 0 to 1.

If the pores within the soil or rock material are interconnected, water can flow through the material and it is said to be "permeable" (Figure 3). Material can be porous without being permeable, but it cannot be permeable without being porous.

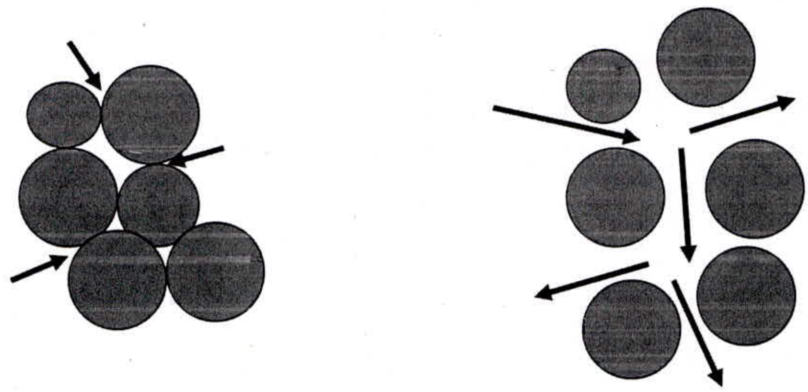

Figure 3: (a) Porous but not permeable (b) Porous and permeable

Moisture content (θ) is defined as the liquid volume within a porous medium per unit volume of the medium. For a fully saturated porous medium, where the pores are completely filled with water, the moisture content is equivalent in magnitude to the porosity. *θ can be expressed as*

$$\theta = Vw/Vt$$

Where, Vw is the volume of water.

If a material's pores are saturated with water, $\theta = n$;

If the pores contain some air, $\theta < n$

Aquifer Storage

In principle, there are two mechanisms by which the water volume in storage in a porous medium may change. Both of these are reversible responses of the aquifer to changes in hydraulic head. The first mechanism, which occurs in unconfined aquifers, is by simple gravity drainage of water from the pore volume. This might occur in the drainage of water near the upper surface of an unconfined aquifer in response to nearby pumping. The second mechanism, which occurs in confined and semi-confined aquifers is a reduction in hydraulic head, which leads to expansion of the water and compaction of the aquifer. As might be anticipated, relatively small amounts of water are typically derived from this second process.

The ratio of the water volume that drains under gravity, V_w, to the total dewatered volume, V_T, is referred to as the specific yield, S_y

$$S_y = \frac{V_w}{V_T} = \frac{V_w}{A \times \Delta h}$$

Where,

S_y = Specific yield, (dimensionless)

V_w = Volume of water drained by gravity from the media, (m^3)

V_T = Total volume, (m^3)

A = Cross-sectional area of the porous media sample, (m^2)

Δh = Drop in the water elevation level in the sample media due to drainage under gravity, (m)

The specific yield is often referred to as the "drainable porosity", because it is the portion that is able to drain under gravity. Since, the total volume of water used per unit volume of a saturated media is the total porosity, *n*, the porosity is a physical upper bound to the magnitude of specific yield in any media sample, i.e., $S_y \leq n$.

In confined and semi-confined aquifers, lesser water volumes may be derived from storage without drainage of the pore volumes by the reduction of the pore water pressure and the resultant compaction of the aquifer matrix. Even though the water pressure is reduced, the buried soil volume must still bear the unchanged overlying load imposed by the soil column. This results in a shift of the burden carried by the water to the soil matrix. In the case of an unconsolidated media, this results in an increase in the intergranular stress – the compressive forces at the point of contact between grains. Most frequently, the grains themselves are assumed to be incompressible, so reductions in pore volume occur through the rearrangement of the matrix grains in response to increases in intergranular stress. This is generally considered to be a reversible process, in that later increases in pore water pressure can cause a return to the original pore volume.

Specific storage (also called the specific storativity), S_s, is defined as the volume of water removed per unit volume of porous media per unit change in hydraulic head. That is:

$$S_s = \frac{S}{b} = \frac{\Delta V_w}{\Delta h\, V_B} = \gamma\beta\theta - \gamma\sigma$$

Where:

S_s = Specific storativity or specific storage, (1/m)

S = Aquifer Storativity, (dimensionless)

b = Saturated aquifer thickness, (m)

ΔV_w = Volume of water, (m^3)

Δh = Change in the hydraulic head, (m)

V_B = Bulk volume of porous medium, (m^3)

γ = Specific weight of water, (N/m^3)

β = Compressibility of water, (m^2/N)

θ = Effective porosity, (dimensionless)

σ = Compressibility of aquifer skeleton, (m^2/N)

Darcy Equation

Henri Darcy established empirically that the flux of water through a permeable formation is proportional to the head difference and inversely proportional to distance.

$$V \propto -\Delta h,$$

$$V \propto 1/\Delta L$$

$$V = -K\,(\Delta h/\Delta L)$$

The constant of proportionality is called the *hydraulic conductivity* (K)

$$Q = VA \;(A = \text{total area})$$

$$Q = -KA\,(dh/dL)$$

Darcy velocity is a fictitious velocity since it assumes that flow occurs across the entire cross-section of the soil sample. Flow actually takes place only through interconnected pore channels.

$$V_S = V_D / n$$

Where V_S is seepage velocity and V_D is the Darcy velocity

Example:

Given: Hydraulic conductivity (K) for the aquifer is 50 m/day and porosity (n) is 0.2. The piezometric head in two wells 1000 m apart is 55 m and 50 m respectively, from a common datum. The average thickness of the aquifer is 30 m, and the average width of aquifer is 5 km

Compute: (a) the rate of flow through the aquifer

(b) the average time of travel from the head of the aquifer to a point 4 km downstream

Solution:

Cross-Sectional area= $30(5)(1000) = 15 \times 10^4\ m^2$

Hydraulic gradient = $(55\text{-}50)/1000 = 5 \times 10^{-3}$

Rate of Flow for K = 50 m/day

$Q = (50\ m/day)\ (75 \times 10^1\ m^2)$

$= 37{,}500\ m^3/day$

Darcy Velocity: $V = Q/A = (37{,}500 m^3/day) / (15 \times 10^4 m^2)$

$= 0.25\ m/day$

Seepage Velocity: $V_s = V/n = (0.25) / (0.2) = 1.25\ m/day$

Time to travel 4 km downstream:

$T = 4(1000m) / (1.25m/day)$

$= 3200$ days or 8.77 years

This example shows that water moves very slowly underground.

Groundwater Flow Equation

To develop mathematical model of almost any system, it is required to formulate general equations. General equations are differential equations that derive from the physical principles governing the process that is to be modelled (Fitts, 2002). In the case of groundwater flow, the physical principles are Darcy's law and mass balance. By combining the mathematical relations describing these principles, a general groundwater flow equation can be derived, which is a partial differential equation.

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

A mathematical model of head ($h\ (x, y, z, t) = \ldots$) must obey this partial differential equation

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Fitts, Charles, R. (2002). Groundwater Science. Academic Press, an imprint of Elsevier cience. ISBN 0-12-257855-4

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Website: http://nptel.ac.in/courses/105103026/module3/lec22/2.html