Flow Through Saturated Soil

Lecture 1

1.0 INTRODUCTION

Study of subsurface flow in saturated soil is so important since about 30% of the world's fresh water resources exist in the form of groundwater. Further, the subsurface water forms a critical input for the sustenance of life and vegetation in arid zones. Due to its importance as a significant source of water supply, various aspects of groundwater dealing with the exploration, development and utilization have been extensively studied by workers from different disciplines, such as geology, geophysics, geochemistry, agricultural engineering, fluid mechanics and civil engineering .

2.0 FORMS OF SUBSURFACE WAYER

Water in the soil mantle is called subsurface water and is considered in two zones (Fig.2.1) :

1- ***Saturated zone***
2- ***Aeration zone***

Saturated zone

This zone, also known as groundwater zone, is the space in which all the pores of the soil are filled with water. The water table forms its upper limit and marks a free surface, i.e. a surface having atmospheric pressure.

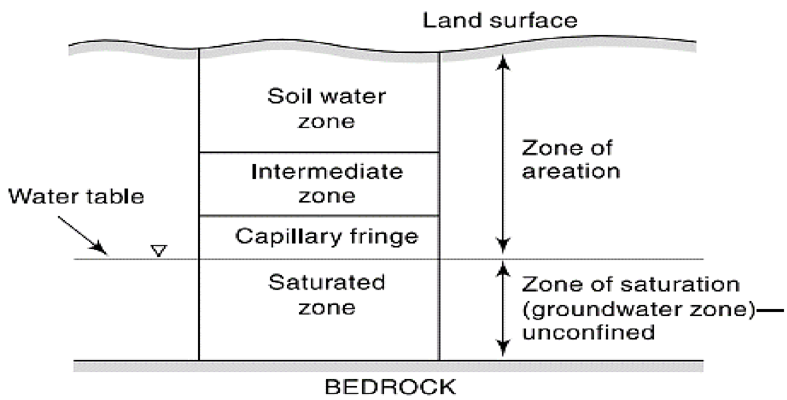

Fig. 2.1 : Classification of Subsurface Water.

Zone of Aeration

In this zone the soil pores are only partially saturated with water. The space between the land surface and the water table marks the extent of this zone. The zone of aeration has three subzones:

CHAPTER 2

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ASS. Prof. Dr. Husham T. Ibrahim

Soil Water Zone

This lies close to the ground surface in the major root band of the vegetation from which the water is lost to the atmosphere by evapotranspiration.

Capillary Fringe

In this the water is held by capillary action. This zone ex- tends from the water table upwards to the limit of the capillary rise.

Intermediate Zone

This lies between the soil water zone and the capillary fringe.

The thickness of the zone of aeration and its constituent subzones depend upon the soil texture and moisture content and vary from region to region. The soil moisture in the zone of aeration is of importance in agricultural practice and irrigation engineering. **The present chapter is concerned only with the saturated zone.**

Saturated Formation

All earth materials, from soils to rocks have pore spaces. Although these pores are completely saturated with water below the water table, from the groundwater utilization aspect only such material through which water moves easily and hence can be extracted with ease are significant.

On this basis the saturated formations are classified into four categories:

1. Aquifer, 2. Aquitard, 3. Aquiclude, and 4. Aquifuge.

Aquifer

An aquifer is a saturated formation of earth material which not only stores water but yields it in sufficient quantity. Thus an aquifer transmits water relatively easily due to its high permeability. Unconsolidated deposits of sand and gravel form good ***aquifers***.

Aquitard

It is a formation through which only seepage is possible and thus the yield is insignificant compared to an aquifer. It is partly permeable. A sandy clay unit is an example of aquitard. Through an ***aquitard*** appreciable quantities of water may leak to an aquifer below it.

Aquiclude

It is a geological formation which is essentially impermeable to the flow of water. It may be considered as closed to water movement even though it may contain large amounts of water due to its high porosity. Clay is an example of an ***aquiclude***.

Aquifuge

It is a geological formation which is neither porous nor permeable. There are no interconnected openings and hence it cannot transmit water. Massive compact rock without any fractures is an ***aquifuge***.

The definitions of ***aquifer***, ***aquitard*** and ***aquiclude*** as above are relative. A formation which may be considered as an aquifer at a place where water is at a premium (e.g. arid zones) may be classified as an ***aquitard*** or even ***aquiclude*** in an area where plenty of water is available.

The availability of groundwater from an aquifer at a place depends upon the rates of withdrawal and replenishment (recharge). Aquifers play the roles of both a transmission conduit and a storage. Aquifers are classified as ***unconfined aquifers*** and ***confined aquifers*** on the basis of their occurrence and field situation. An unconfined aquifer (also known as water table aquifer) is one in which a free water surface, i.e. a water table exists (Fig. 2.2). Only the saturated zone of this aquifer is of importance in groundwater studies. Recharge of this aquifer takes place through infiltration of precipitation from the ground surface. A well driven into an unconfined aquifer will indicate a static water level corresponding to the water table level at that location.

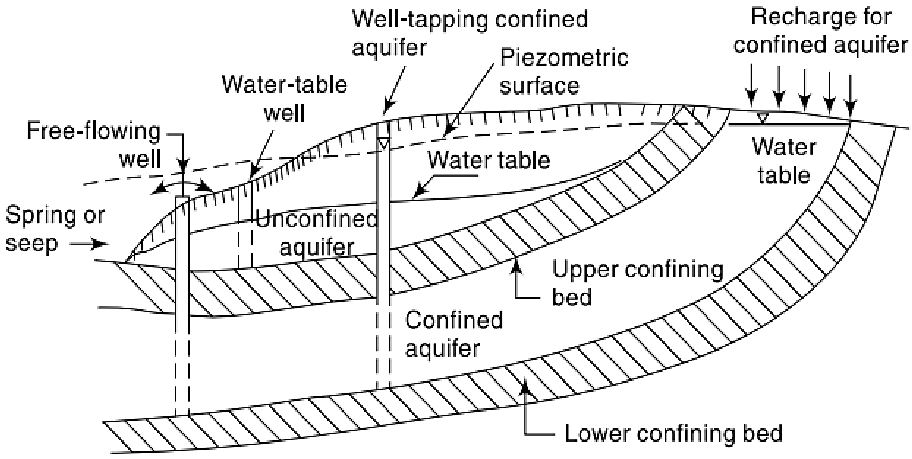

Fig. 2.2 Confined and Unconfined Acquifers

A confined aquifer, also known as artesian aquifer, is an aquifer which is confined between two impervious beds such as aquicludes or aquifuges (Fig. 2.2). Recharge of this aquifer takes place only in the area where it is exposed at the ground surface. The water in the confined aquifer will be under pressure and hence the piezometric level will be much higher than the top level of the aquifer. At some locations: the piezomet- ric level can attain a level higher than the land surface and a well driven into the aquifer at such a location will flow freely without the aid of any pump. A confined aquifer is called a leaky aquifer if either or both of its confining beds are aquitards. A confined aquifer is called ***a leaky aquifer*** if either or both of its confining beds are aquitards.

3.0 WATER TABLE

A water table is the free water surface in an unconfined aquifer. The static level of a well penetrating an unconfmed aquifer indicates the level of the water table at that point. The water table is constantly in motion adjusting its surface to achieve a balance between the recharge and outflow from the subsurface storage. Fluctuations in the water level in a dug well during various seasons of the year, lowering of the groundwater table in a region due to heavy pumping of the wells and the rise in the water table of an irrigated area with poor drainage, are some common examples of the fluctuation of the water table. In a general sense, the water table follows the topographic features of the surface. If the water table intersects the land surface the groundwater comes out to the surface in the form of springs or seepage.

Sometimes a lens or localised patch of impervious stratum can occur inside an unconfined aquifer in such a way that it retains a water table above the general water table (Fig. 2.3). Such a water table retained around the impervious material is known as perched water table. Usually the perched water table is of limited extent and the yield from such a situation is very small. In groundwater exploration, a perched water table is quite often confused with a general water table.

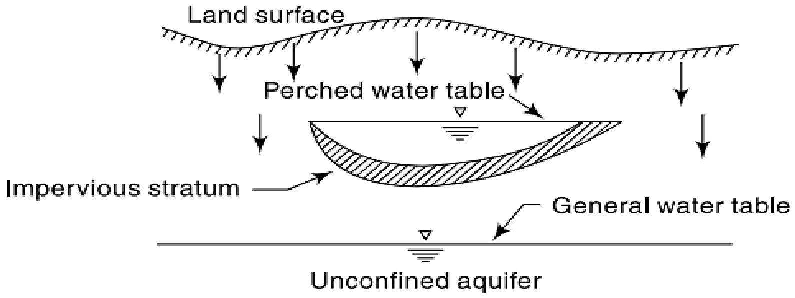

Fig. 2.3 : Perched Water Table

The position of the water table relative to the water level in a stream determines whether the stream contributes water to the groundwater storage or the other way about. If the bed of the stream is below the groundwater table, during periods of low flows in the stream, the water surface may go down below the general water table elevation and the groundwater contributes to the flow in the stream. Such streams which receive groundwater flow are called effluent streams (Fig. 2.4 (a)). Perennial rivers and streams are of this kind. If, however, the water table is below the bed of the stream, the stream-water percolates to the groundwater storage and a hump is formed in the groundwater table (Fig. 2.4 (b)). Such streams which contribute to the groundwater are known as influent streams. Intermittent rivers and streams which go dry during long periods of dry spell (i.e. no rain periods) are of this kind.

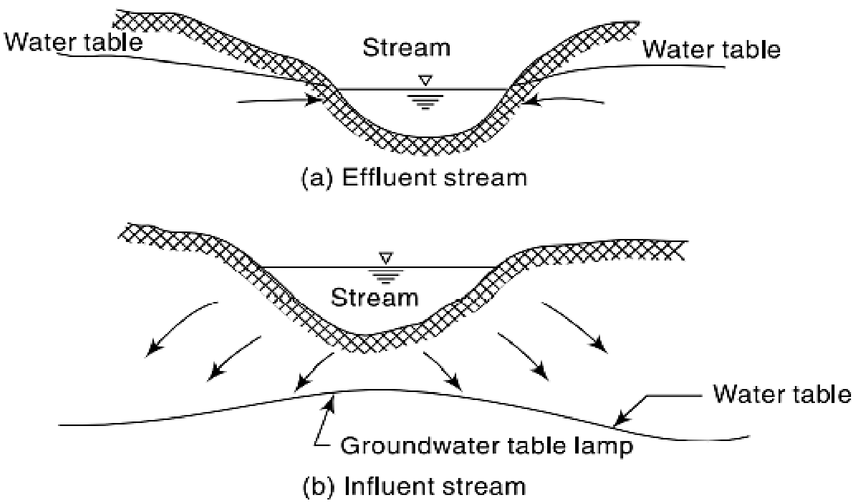

Fig. 2.4: Effluent and Influent Streams

4.0 AQUIFER PROPERITIES

The important properties of an aquifer are its capacity to release the water held in its pores and its ability to transmit the flow easily. These properties essentially depend upon the composition of the aquifer.

Porosıty

The amount of pore space per unit volume of the aquifer material is called porosity. It is expressed as:

$$n = \frac{V_v}{V_0} \quad \text{------- (2.1)}$$

where n = porosity, V_v = volume of voids and V_0 = volume of the porous medium. In an unconsolidated material the size distribution, packing and shape of particles determine the porosity. In hard rocks the porosity is dependent on the extent, spacing and the pattern of fracturing or on the nature of solution channels. In qualitative terms porosity greater than 20% is considered as large, between 5 and 20% as medium and less than 5% as small.

Specific Yield

While porosity gives a measure of the water-storage capability of a formation, not all the water held in the pores is available for extraction by pumping or draining by gravity. The pores hold back some water by molecular attraction and surface tension. The actual volume of water that can be extracted by the force of gravity from a unit volume of aquifer material is known as the ***specific yield (S_y)***. The fraction of water held back in the aquifer is known as ***specific retention***, (S_r). Thus porosity :

$$n = S_y + S_r \quad \text{----(2.2)}$$

The representative values of porosity and specific yield of some common earth materials are given in Table 2.1

Table 2.1 : 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

It is seen from Table 2.1 that although both clay and sand have high porosity the specific yield of clay is very small compared to that of sand.

5.0 DARCY's LAW

In 1856 Henry Darcy, a French hydraulic engineer, on the basis of his experimental findings proposed a law relating the velocity of flow in a porous medium. This law, known as *Darcy's law,* can be expressed as

$$V = Ki \qquad \text{-----(2.3)}$$

where V = Apparent velocity of seepage = Q/A in which Q = discharge and A = cross-sectional area of the porous medium. V is sometimes also known as discharge velocity. $i = -\dfrac{dh}{ds}$ = hydraulic gradient, in which h = piezometric head and s = distance measured in the general flow direction; the negative sign emphasizes that the piezometric head drops in the direction of flow. K = a coefficient, called *coefficient of permeability,* having the units of velocity.

The discharge Q can be expressed as

$$Q = K\,i\,A$$

$$= K\,A\left(-\frac{\Delta H}{\Delta s}\right) \qquad \text{----(2.3 a)}$$

where $(-\Delta H)$ is the drop in the hydraulic grade line in a length Δs of the porous medium.

Darcy's law is a particular case of the general viscous fluid flow. It has been shown valid for laminar flows only. For practical purposes, the limit of the validity of Darcy's law can be taken as Reynolds number of value unity, i.e.

$$\mathbf{Re} = \frac{V\,d_a}{\nu} = 1 \qquad \text{------(2.4)}$$

where **Re** = Reynolds number

d_a = representative particle size, usually $d_a = d_{10}$ where d_{10} represents a size such that 10% of the aquifer material is of smaller size.

ν = kinematic viscosity of water

Except for flow in fissures and caverns, to a large extent groundwater flow in nature obeys Darcy's law. Further, there is no known lower limit for the applicability of Darcy's law.

It may be noted that the *apparent velocity* V used in Darcy's law is not the actual velocity of flow through the pores. Owing to 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 the porous media is expressed as

$$v_a = \frac{V}{n} \qquad \text{-------(2.5)}$$

where n = porosity. The bulk pore velocity v_a is the velocity that is obtained by tracking a tracer added to the groundwater.

6.0 COFFICIENT OF PERMEABLITY

The coefficient of permeability, also designated as *hydraulic conductivity* reflects the combined effects of the porous medium and fluid properties. From an analogy of laminar flow through a conduit (*Hagen-Poiseuille flow*) the coefficient of permeability K can be expressed as

$$K = Cd_m^2 \frac{\gamma}{\mu} \qquad \text{------(2.6)}$$

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 which depends on the porosity, packing, shape of grains and grain-size distribution of the porous medium. Thus for a given porous material

$$K \propto \frac{1}{\nu}$$

where ν = kinematic viscosity = μ/ρ = f(temperature). The laboratory or *standard value* of the coefficient of permeability (K_s) is taken as that for pure water at a standard temperature of 20° C. The value of K_t, the coefficient of permeability at any temperature t can be converted to K_s by the relation

$$K_s = K_t(\nu_t/\nu_s) \qquad \text{------(2.7)}$$

where ν_s and ν_t represent the kinematic viscosity values at 20° C and t°C respectively.

The coefficient of permeability is often considered in two components, one reflecting the properties of the medium only and the other incorporating the fluid properties. Thus, referring to Eq. (2.6) , a term K_0 is defined as

$$K = K_0 \frac{\gamma}{\mu} = K_0 \frac{g}{\nu} \qquad \text{-----(2.8)}$$

where $K_0 = C\,d_m^2$. The parameter K_0 is called specific or *intrinsic permeability* which is a function of the medium only. Note that K_0 has dimensions of $[L^2]$. It is expressed in units of cm^2 or m^2 or in darcys where 1 darcy = 9.87×10^{-13} m^2. Where more than one fluid is involved in porous media flow or when there is considerable temperature variation, the coefficient K_0 is useful. However, in groundwater flow problems, the temperature variations are rather small and as such the coefficient of permeability K is more convenient to use. The common units of K are m/day or cm/s. The conversion factor for these two are

1 m/day = 0.0011574 cm/s

or 1 cm/s = 864.0 m/day

Some typical values of coefficient of permeability of some porous media are given in Table 2.2.

Table 2.2 : Representative Values of the Permeability Coefficient

No. Material	K (cm/s)	K_0 (darcys)
A. Granular material		
1. Clean gravel	1–100	10^3–10^5
2. Clean coarse sand	0.010–1.00	10–10^3
3. Mixed sand	0.005–0.01	5–10
4. Fine sand	0.001–0.05	1–50
5. Silty sand	$1 \times 10^{-4} - 2 \times 10^{-3}$	0.1–2
6. Silt	$1 \times 10^{-5} - 5 \times 10^{-4}$	0.01–0.5
7. Clay	$< 10^{-6}$	$< 10^{-3}$
B. Consolidated material		
1. Sandstone	$10^{-6} - 10^{-3}$	$10^{-3} - 1.0$
2. Carbonate rock with secondary porosity	$10^{-5} - 10^{-3}$	$10^{-2} - 1.0$
3. Shale	10^{-10}	10^{-7}
4. Fractured and weathered rock (aquifers)	$10^{-6} - 10^{-3}$	$10^{-3} - 1.0$

At 20° C, for water, $\nu = 0.01$ cm^2/s and substituting in Eq. (2.8)

$$K_0 \text{ [darcys]} = 10^3 K \text{ [cm/s] at 20°C}$$

Consider an aquifer of unit width and thickness B, (i.e. depth of a fully saturated zone). The discharge through this aquifer under a unit hydraulic gradient is

$$T = KB \qquad \text{---- (2.9)}$$

This discharge is termed *transmissibility, T* and has the dimensions of $[L^2/T]$. Its units are m^2/s or litres per day/metre width (l pd/m). Typical values of T lie in the range 1×10^6 l pd/m to 1×10^4 l pd/m. A well with a value of $T = 1 \times 10^5$ l pd/m is considered satisfactory for irrigation purposes.

The coefficient of permeability is determined in the laboratory by a *permeameter.* For coarse-grained soils a *constant-head permeameter* is used. In this the discharge of water percolating under a constant head difference (ΔH) through a sample of porous material of cross-sectional area A and length L is determined. The coefficient of permeability at the temperature of the experiment is found as

$$K = \frac{Q}{A} \frac{1}{(\Delta H/L)}$$

Under field conditions, permeability of an aquifer is determined by conducting pumping tests in a well. One of the many tests available for this purpose consists of pumping out water from a well at a uniform rate till steady state is reached. Knowing the steady-state drawdown and the discharge-rate, transmissibility can be calculated. Information about the thickness of the saturation zone leads one to calculate the permeability. Injection of a tracer, such as a dye and finding its velocity of travel is another way of determining the permeability under field conditions.