1. Physical and Engineering Properties of Rocks

1.1 Introduction

The performance of soil and rock under the action of load, water, temperature and tectonics of earth crust depends upon physical and mechanical strength properties of those materials. There are several classifications for rocks properties, of these the physical properties (***Index properties***) and the mechanical (or engineering) properties or the strength properties.

1.2 Physical Properties

The physical properties of rocks affecting design and construction in rocks are:
1-Bulk density
2- Unit weight
3- Specific gravity
4- Porosity and void ratio
5- Dry and saturated unit weight
6- Moisture content
7- Degree of saturation

1.2.1 Bulk Density (ρ)

It is the ratio between rock mass (***M***) and its volume (***V***) which is the average density and also known as the bulk density. Its units are gm/cm ^{3} or 1000 kg/m^3. For most rocks near earth surface have average densities between (1.5 – 3) gm/cm^3.

$$\rho = \frac{M}{V} \qquad (1.1)$$

1.2.2 Unit Weight (γ)

It is also known as weight density **(γ)**, it is the ratio of rock weight to its volume and its unit is N / m^3.

$$\gamma = \frac{W}{V} = \frac{N}{m^3} \qquad (1.2)$$

Where $W = M \cdot g$; so (1.3)

$$\gamma = \frac{M \cdot g}{V} = \left(\frac{M}{V}\right) \cdot g = \rho \cdot g \quad ; \quad \longrightarrow \quad \gamma = \rho \cdot g \qquad (1.4)$$

where g is the acceleration = 9.8 m / s^2
The unit weight of water is:
$\gamma_w = \rho_w \cdot g$ = (1000kg/ m^3) (9.8 m / s^2)=9.8 kN/m^3

1.2.3 Specific Gravity (*G*)

It is an essential property in engineering projects, represents the ratio of rock density (or its unit weight) to water density (or its unit weight), so that ***G*** has no units.

$$G = \frac{\rho}{\rho_w} \quad ; \qquad (1.5)$$

$$G = \frac{\rho \cdot g}{\rho_w \cdot g} \quad ; \quad \longrightarrow \quad G = \frac{\gamma}{\gamma_w} \qquad (1.6)$$

EXAMPLE 1.1: Calculate the density in gm/cm ^{3} of a rock with unit weight 27.6 kN/m^3.
$\gamma = \rho \cdot g$
27.6 * 10^3 N/m^3 = ρ . (9.8 m / S^2)
ρ = 2.82 gm/cm ^{3}

EXAMPLE 1.2: A block of rock with edge length 85.5cm, 79.0cm, 43.8cm has a mass of 953 kg. Find the specific gravity of the rock?
V = (0.855m) (0.79m) (0.438m) = 0.2958 m^3

$$\rho = \frac{M}{V} = \frac{953\text{kg}}{0.2958\ \text{m}^3} = 3222\ \text{kg / m}^3$$

$$G = \frac{\rho}{\rho_w} = \frac{3222\ \text{kg / m}^3}{1000\ \text{kg / m}^3} = 3.22$$

1.2.4 Porosity (*n*)

It is the percentage ratio of the volume of voids to the total volume, expressed either as percentage or as a fraction.

$$n = \frac{V_v}{V} * 100 \qquad (1.7)$$

$V = V_v + V_s$ (1.8)

where V_v = volume of voids; V_s = volume of grain (mineral)

The density ρ_g of the grains is given by the following:

$$\rho_g = \frac{M_{grain}}{V_{grain}} = \frac{M_g}{V_g} = \frac{M_s}{V_s} = \rho_s \qquad (1.9)$$

$$V_g = V - V_v \quad ; \quad V_v = nV \longrightarrow \quad V_g = V - nV$$

$$V_g = (1 - n)V \qquad (1.10)$$

As $V_g = M_g / \rho_g$; substituting in the later equation:

$M_g / \rho_g = (1-n)V$; $\rho_g = M / (1-n) V$

$\rho_g = \rho / 1-n$; and (1.11)

$\gamma_g = \gamma / 1-n$ (1.12)

The same conclusion can be reached using the following relations:

$\rho_g = M_g / V_g$ and $V_g = V - V_v$ so $\rho_g = M_g / V - V_v$ and $V_v = nV$

$\rho_g = M_g / V - nV$ and $M_g = M$ so $\rho_g = M / V(1-n)$ $\rho_g = \rho / 1-n$

Void Ratio (*e*)

It is the ratio of volume of voids to the volume of solid, expressed by the following relation:

$e = V_v / V_s$; (1.13)

$V = V_v + V_s$ so $V_s = V - V_v$

$e = V_v / V - V_v$; and $n = V_v / V$ so $V_v = nV$

$e = nV / V - nV = nV / V(1-n)$ so $e = n / 1-n$ (1.14)

EXAMPLE 1.3: A 0.885 m^3 block of sandstone has a mass of 1752 kg. When the bock is crushed just sufficiently to close all the pores , which are empty, the volume of the rock becomes 0.584 m^3. Find: (a) the porosity of the sandstone and (b) the density of the grains. (Assume that the density of the grains is not changed in the crushing process).

(a) The original block has a volume V= 0.885 m^3. The volume of the crushed rock must equal the volume of all the grains in the original block, since the crushed rock has zero pore volume. That is , in the original block, V_{grains} = 0.584 m^3. Therefore, the volume of the pores in the original block is:

$$V=V_{pores}+V_{grains}$$

0.885 m^3= V_{pores} + 0.584 m^3 V_{pores}=0.301 m^3

Now calculate the porosity :

$n= V_v / V$ n= (0.301 m^3) / (0.885 m^3) ; n= 0.340 (34%)

(b) Since the pores are empty, the mass of the crushed rock is the same as that of the original block, 1752 kg. The volume of the crushed rock is 0.584 m^3. But the crushed rock is entirely grains, and therefore:

$\rho_g=M_{grain} / V_{grain}$; ρ_g = (1752 kg) / (0.584 m^3)= 3000kg / m^3 = 3.00 gm/cm^3

1.2.5 Dry and Saturated Unit Weights

The pores of in situ rock (rock as found in the earth, undisturbed by human activity) may be filled with gas or liquid. The densities of gases found in rocks are much less than the densities of the grains of the rocks. This means that, it is safe to ignore the gas trapped in the pores to the total weight of a rock sample.

A similar statement cannot be made for liquids. The densities of liquids commonly found in the pores, although less than the densities of the grains, are not very much less. If the porosity of the rock is large enough, and a significant fraction of the pores contain liquid , then the weight (or mass) of the liquid is likely to be an appreciable part of the total weight (or mass) of the rock.

If all the pores in a rock sample are completely filled with liquid , then the rock is said ***saturated***. If all the pores are empty, then the rock is said to be ***dry***. Saturated rock can be rendered by heating the rock in an oven, at sufficiently high temperature the liquid vaporizes and the vapor is driven out of the rock.

An important relation is that between the unit weight of a saturated rock sample γ_{sat} , the unit weight of the same sample when dry γ_{dry} , and the unit weight of the liquid occupying the pores of the saturated sample γ_L .

It may be tempting to write $\gamma_{sat} = \gamma_{dry} + \gamma_L$, but this is not correct because of the different volumes involved.

To obtain the actual relation, note that the weight W_{sat} of the saturated rock sample is just the sum of the dry weight W_{dry} and the weight W_L of the liquid in the saturated rock:

$$W_{sat} = W_{dry} + W_L$$

The volume V of the rock sample is the same whether it is dry or saturated. Dividing the preceding equation by V gives:

$W_{sat} / V = W_{dry} / V + W_L / V$;
$\gamma_{sat} = \gamma_{dry} + W_L / V$; (1.15)

and $\gamma_L = W_L / V_L$; (1.16)
where V_L is the volume of liquid with weight W_L. But, since the liquid fills all the pores;
$V_L = V_{pores}$ (1.17)
$V_L = nV$ (1.18)
Solve this last equation for **V** and substitute into Eq.(8.15). Then use the definition of γ_L given above to obtain:
$\gamma_{sat} = \gamma_{dry} + n\ (W_L / V_L)$;
$\gamma_{sat} = \gamma_{dry} + n\ .\ \gamma_L$; or $\gamma_{sat} = \gamma_{dry} + n\ .\ \gamma_w$ (1.19)
A similar relation holds between the mass densities :
$\rho_{sat} = \rho_{dry} + n\ .\ \rho_L$; or $\rho_{sat} = \rho_{dry} + n\ .\ \rho_w$ (1.20)

EXAMPLE 1.4 : A test cylinder of rock has a diameter of 12.6 cm and a length of 14.0 cm. When dry its weight is 50.3 N. When saturated with mercury, the weight of the sample is 62.8 N. The specific gravity of mercury is 13.6. Find the porosity of the rock?

The volume of the rock sample is:
$V = \pi D^2 L/4 = \pi(0.126\text{m})^2\,(0.140)/4 = 1.746*10^{-3}\ \text{m}^3$
Therefore, the dry and saturated unit weights are
$\gamma_{dry} = W_{dry} / V = (50.3\ \text{N}) / (1.746*10^{-3}\ \text{m}^3) = 28.81\ \text{kN / m}^3$
$\gamma_{sat} = W_{sat} / V = (62.8\ \text{N}) / (1.746*10^{-3}\ \text{m}^3) = 35.97\ \text{kN / m}^3$
The unit weight of the liquid mercury γ_L follows from equation:
$\gamma_L = G_L\, \gamma_w = (13.6)\ (9.8\ \text{kN / m}^3) = 133.3\ \text{kN / m}^3$
Note that the unit weight of the mercury is grater than the unit weight of the rock, whether dry or saturated. Now solve for the porosity n.
$\gamma_{sat} = \gamma_{dry} + n\ .\ \gamma_L$
$35.97\ \text{kN / m}^3 = 28.81\ \text{kN / m}^3 + n\ (133.3\ \text{kN / m}^3)$
$n = 0.0537\ (5.37\%)$

1.2.6 Multimineral Rocks

Suppose that the porosity of a particular rock specimen is to be found by measuring the bulk density ρ, the grain density ρ_g and then applying Eq.(1.11):
$\rho_g = \rho / 1\text{-}n$
The bulk density is easy to measure. What about grain density ? If all of the grains in the rock are of the same mineral, and the density of the mineral as it occurs in nature is known (many have been measured in the laboratory), then the grain density simply equals the density of that mineral.

For a rock that contains several minerals, the value of ρ_g to use in above equation is the average of the densities of the individual minerals present. As an example, consider a rock made up of three minerals , the densities of the minerals grains being denoted by ρ_1, ρ_2 and ρ_3. The bulk grain density ρ_g will not, in general, be simply $1/3\ (\rho_1 + \rho_2 + \rho_3)$, because the minerals may be present in different amounts. A ***weighted average*** must be used, the precise nature of which must now be deduced.

Let the total mass of all the grains in the rock sample be M_g and the total volume of all the grains V_g, so

$M_g = \rho_g V_g$

If M_1 be the total mass and V_1 total volume of mineral ***1*** in the rock, with similar notation for the other two minerals present (assuming that the rock contains three minerals), then since

$M_g = M_1 + M_2 + M_3$ it follows that:

$\rho_g V_g = \rho_1 V_1 + \rho_2 V_2 + \rho_3 V_3$;

$\rho_g = \rho_1 (V_1 / V_g) + \rho_2 (V_2/ V_g) + \rho_3 (V_3/ V_g)$;

and if : $c_1 = V_1/ V_g$, $c_2 = V_2 / V_g$ and $c_3 = V_3 / V_g$ so that : (1.21)

$$c_1 + c_2 + c_3 = 1 \quad (1.22)$$

$$\rho_g = c_1 \rho_1 + c_2 \rho_2 + c_3 \rho_3 \quad (1.23)$$

In similar way, for the specific gravity:

$$G = c_1 G_1 + c_2 G_2 + c_3 G_3 \quad ; \quad G = \Sigma_i c_i G_i \quad (1.24)$$

EXAMPLE 1.5 : A shale consists of 34.1% chlorite and 65.9% pyrite, and has a porosity of 38.8% knowing that the density of the chlorite is 2.8 gm/cm^3 and for pyrite is 5.05 gm/cm^3. Find the bulk density of the shale?

First, find the bulk grain density:

$\rho_g = c_1 \rho_1 + c_2 \rho_2$ =(0.341)(2.80 gm/cm^3) + (0.659)(5.05 gm/cm^3)=4.283 gm/cm^3

$\rho_g = \rho / 1\text{-}n$

$\rho = \rho_g (1\text{-}n)$=(4.283 gm/cm^3)(1- 0.385) = 2.62 gm/cm^3

1.2.7 Water Content (W_c)

It is defined as the ratio of the weight of water W_w to the weight of solids W_s and it is dimensionless, expressed either as percentage or as a fraction:

$$W_c = W_w / W_s = (W - W_s) / W_s \quad (1.25)$$

Where W is the total weight of the rock sample and W_s is the weight of the solids. Note, some texts refers to the total weight instead of the weight of the solids.

1.2.8 Degree of Saturation (S)

It expresses the degree to which the voids are filled with water, so it is the ratio of the total volume of all water, which is found in the voids, in the soil sample (V_w) to the total volume of voids (V_v), and has no units, expressed either as percentage or as a fraction. Rocks with high porosity have higher degree of saturation which are found beneath groundwater table.

$\boldsymbol{S = V_w / V_v}$ (1.26)

1.2.9 Other Relations among Different Physical Properties

There are many Mathematical relations related physical properties with each other, of these re the following:

$e = n / 1-n$;

$n = e / (1+e)$; (1.27)

$n = W_c \,.\, G / (1+ W_c.G)$; (1.28)

$\gamma_{dry} = \gamma_{wet} / (1 + W_c)$; (1.29)

$\gamma_{dry} = G \ \gamma_w / (1 + e)$; (1.30)

$\gamma_{dry} = G \, \gamma_w \, (1 - n)$; (1.31)

$\gamma_{sat} = (G + e) \, \gamma_w \ / (1 + e)$ (1.32)

EXAMPLE 8.6 : A sandstone core with specific gravity 2.63 composed of quartz and feldspar grains with calcite cement is 82 mm in diameter and 169 mm long. On saturation in water, its weight is 21.42 N; after oven drying, its weight is 20.31 N. Calculate: (a)its porosity (b) its dry unit weight (c) its wet unit weight?

$W_c = (W_{sat} - W_{dry}) / W_{sat} = (21.42 \text{ N} - 20.31 \text{ N}) / 21.42 \text{ N} = 0.0518 \ (5.18\%)$

$n = W_c \,.\, G / (1+ W_c\,.G) = 0.0518 * 2.63 / (1+ 0.0518 * 2.63) = 0.12 \ (12\%)$

$\gamma_{dry} = G \, \gamma_w \, (1 - n) = 2.63 * (1000 \text{ kg} / \text{m}^3 * 9.8) \text{ N} / \text{m}^3 \, (1 - 0.12)$

$\gamma_{dry} = 22.68 \text{ k N} / \text{m}^3$

$\gamma_{dry} = \gamma_{wet} / (1 + W_c)$

$22.68 \text{ kN} / \text{m}^3 = \gamma_{wet} / (1 + 0.0518)$

$\gamma_{wet} = 23.85 \text{ kN} / \text{m}^3$

1.3 Rock Deformation and Mechanical Properties

Rocks in their natural state are fractured, inhomogeneous, anisotropic and discontinuous. The construction design in rocks requires the knowledge of the resulted deformations which can be achieved by knowing their mechanical properties (strengths) in order to perform the suitable design.

1.3.1 Rock Deformation

Deformation in rocks means the change in size and shape of the rock sample induced by the applied force even though the rock sample does not break.

Let us start with the unconfined compression and tensile test. The arrangement of the applied forces for the two tests is called a uniaxial load because the forces are parallel to the axis of the cylinder. Figure 1.1 shows two loadings, together with the rock cylinder when no load (***F***=0). Figure (1.1a) is of a test cylinder of rock ***at rest*** under ***no load***, that is, there are no forces being applied to the cylinder. In this situation, the length of the cylinder is given the symbol L_0 , the diameter D_0 and the cross-sectional area A_0.

In Fig. (1.1b) a uniaxial tensile load is applied. This means that equal and oppositely directed forces F are simultaneously applied perpendicularly to the ends of the cylinder, the forces being directed as to tend to pull the cylinder apart. This load is unconfined since no forces are applied to the sides of the cylinder. Thus the cylinder becomes longer and thinner than the no-load cylinder. It is assumed that the forces are not so large as to bring the cylinder close to rupture.

In Fig. (1.1c) the forces are reversed in direction relative to above Fig. (1.1 b). In this case, the forces tend to compress the cylinder and therefore this situation is described as a ***uniaxial compressive load***. Under such load, the cylinder of rock becomes shorter and thicker than the no-load cylinder.

For both the tensile and compressive load, the length of the cylinder under the load is given the symbol L, the diameter, D and the cross-sectional area, A.

The force divided by area equals ***stress***. When the forces are first applied, the cross-sectional area of the cylinder is A_0. But the cylinder eventually assumes a different cross-sectional area A (it takes time for the cylinder to deform). Which area should be used to define the stress? Either the ***initial*** are A_0 or the ***final*** area A can be used, but the same choice must be maintained throughout a calculation. If the no-load area A_0 is selected , then the stress so calculated is called the

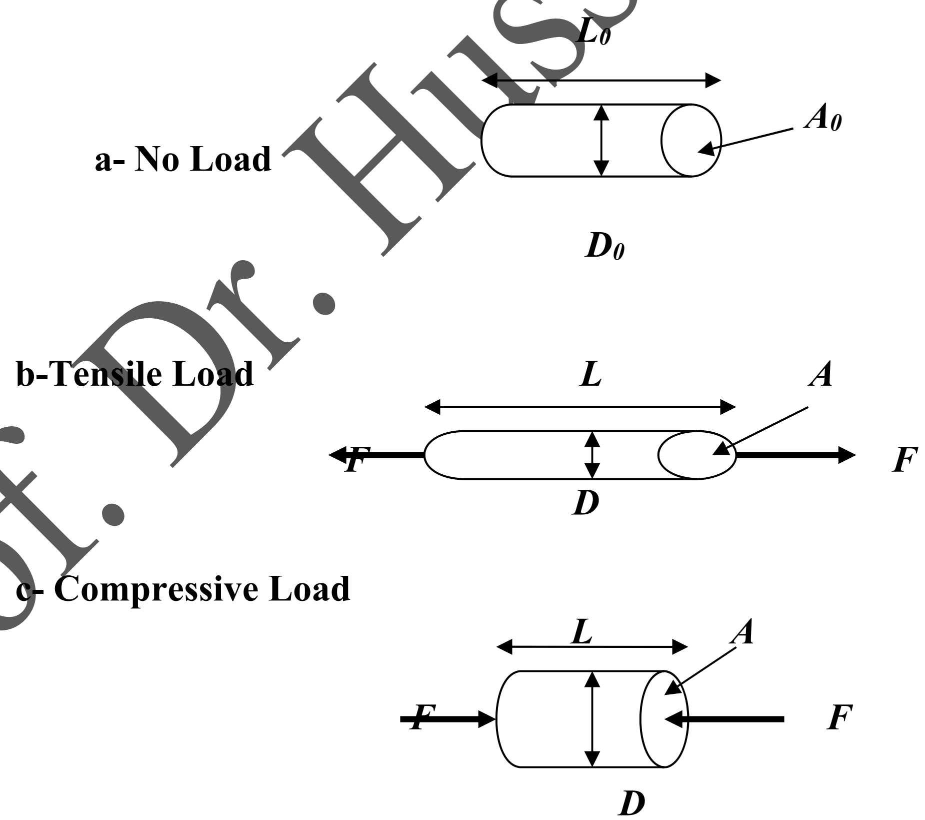


Fig. (1.1). Rock cylinder subjected to tensile and compressive loads.

engineering stress; if the area ***A*** under the load is used , then the stress calculated is called the ***true stress***. That is, the stress **σ** is defined as:

For engineering stress $\sigma_{eng} = F / A_0$; (1.33)
For true stress $\sigma_T = F / A$ (1.34)

As a measure of the rock deformation, the axial strain ϵ_L is defined by:
$\epsilon_L = (L - L_0) / L_0 = \Delta L / L_0$ (1.35)
Note that the true strain to be used with true stress F / A is not $\Delta L / L$ but $\ln (\Delta L / L)$. (In this text, we will work only with engineering stress and strain).

8.3.1.1 Hooke's Law

It is found by actual experiments that , as long as the stress is neither too large nor too small, the stress is directly proportional to the strain, that is ***Hooke's Law***:
$\sigma = E \epsilon$ (1.36)
where **E** is known both as ***Young's modulus*** and as the ***modulus of elasticity***. Since **σ** has units of **Pa** and **ϵ** has no units, and also ***E*** has units of **Pa.**

The axial strain is negative for a compressive load with $L < L_0$ or positive with $L > L_0$. Therefore, for the stress:
$\sigma = \pm F / A_0$ (1.37)
the plus **(+)** sign is used for a tensile load , and the minus (-) sign for a compressive load.

As above Fig. (8.1) suggest a cylinder of rock gets thinner under an axial tensile load and thicker under an axial compressive load. This lateral, or transverse, deformation is expressed by the transverse strain ϵ_T, which is defined in strict analogy with the axial strain ϵ_L. In the below equation then, D_0 is the no-load diameter and D is the diameter under the load, so that:
$\epsilon_T = (D - D_0) / D_0$ (1.38)
Like the axial strain, the transverse strain is dimensionless.

It turns out that, in the elastic region, the range of values of applied stress in which the relation ($\sigma = E \epsilon$) is valid, defined ***by Poisson's ratio***:
$\mu = - \epsilon_T / \epsilon_L$ (1.39)
As both of the strains ϵ_T and ϵ_L are dimensionless, the quantity μ is dimensionless too.

Why the minus sign in the above equation (1.39)? Under an axial tensile load $\epsilon_L > 0$ since $L > L_0$, but $\epsilon_T < 0$ since $D < D_0$ (the cylinder gets longer but thinner). With the minus sign , the above equation yields $\mu > 0$. With an axial compressive load, $\epsilon_L < 0$ since $L < L_0$ and $\epsilon_T > 0$ since $D > D_0$. But again the above equation gives $\mu > 0$. In short, the minus sign in the above equation ensures that Poisson's ratio is a positive quantity ranges between 0.0 and 0.5 ($0.0 < \mu < 0.5$). Its values for rocks are ranging between 0.2 and 0.3.

1.3.1.2 Volumetric Strain

Since the diameter and length of the test cylinder of rock both change under an axial load, it is likely that the volume of the cylinder changes also under the load. For a circular cylinder, the no-load volume V_0 is:

$$V_0 = (1/4)\, \pi\, D_0^2\, L_0 \qquad (1.40)$$

The volume V of the cylinder under the applied load is:

$$V = (1/4)\, \pi\, D^2\, L \qquad (1.41)$$

The later equation assumes that the cylinder of rock remains a cylinder under the load; this is observed to be the case as long as the applied stress does not approach the strength of the rock. To express V in terms of V_0, and therefore be able to compare the volumes under no-load and under a load, write L in terms of L_0 through the axial strain, and D in terms of D_0 through the transverse strain.

And as equations (1.35) and (1.38) can be written as:

$$L = L_0\, (1 + \epsilon_L) \quad ; \quad D = D_0\, (1 + \epsilon_T)$$

Substituting these two equations into Eq.(1.41) and invoking Eq.(1.40); yields

$$V - V_0 / V_0 = \Delta V / V_0 = \epsilon_L + 2\epsilon_T \qquad (1.42)$$

Using Eq.(8.39) $\mu = - \epsilon_T / \epsilon_L$

So $\Delta V / V_0 = \epsilon_L\, (1 - 2\mu)$ (1.43)

The quantity on the left of the later equation looks like a volume strain; this fractional change in volume is called the ***volumetric strain***, or the ***dilatation***.

Now suppose that a rock is ***incompressible***. The volume of a test cylinder of an incompressible rock does not change under a load, so that $V = V_0$. This does not mean that the length and diameter do not change. However, since the volumetric change is zero, so the above equation reduces to:

$$0 = \epsilon_L\, (1 - 2\mu)$$

Since $\epsilon_L \neq 0$ then $0 = (1 - 2\mu)$ so that $\mu = 1/2$

However , no material is truly incompressible, so $\mu < 1/2$. Also lateral strains are not zero in unconfined uniaxial loading, so $\mu = 0$ does not occur. The range of values of Poisson' s ratio for ordinary rocks is $(0.0 < \mu < 1/2)$ where low values for very solid rocks and higher values for over saturated rocks.

8.3.1.3 Elastic Moduli

Several moduli are available, each one expresses a deformation in rocks under a specific mode of loading. Of these moduli we mention the following:

1- Young's Modulus or Modulus of Elasticity (E): It is stress to strain ratio under simple tension or compression and its unit is Pascal**,** Pa (=N/m^2).

$$E = \sigma / \epsilon = (F/A) / \epsilon \qquad (1.44)$$

2- Poisson's ratio (μ): It is a measure of geometric change of shape and it is unitless.

$\mu = (\Delta D/D) / (\Delta L /L)$ (1.45)

3- Shear Modulus or Modulus of Rigidity (*G*): It is stress to strain measure of simple shear, simply called ***rigidity*** and its unit is Pascal, Pa (N/m^2).

G*=shear stress / shear strain = *(F/A) / tan Φ (1.46)

4- **Bulk Modulus or incompressibility (*k*):** It is a measure of stress to strain under simple hydrostatic pressure and its unit is Pascal, Pa (=N/m^2). It is the reverse of ***compressibility*** **(K)**.

k* = volume stress / volume strain = *P* (pressure)/ *(ΔV / V) (1.47)

$\sigma = k\ (\Delta V /V)$; (1.48)

$\Delta V = V - V_0$

$K = 1 / k$ (1.49)

Other relations between the above mentioned moduli are:

$G = E / 2(1 + \mu)$; (1.50)

$k = E / 3\ (1 - 2\ \mu)$ (1.51)

EXAMPLE 1.7: An axial compressive load of 37.6 kN is applied to A cylinder of rock with no-load length of 12.6 cm and a no-load diameter of 4.83 cm. For the cylinder of rock, load Young's Modulus equals 35.0 GPa. Find the change in length of the cylinder under the load?

$\sigma = F / (\pi D_0^2 / 4)$

$\sigma = 37.6*10^3$ N / (π* 0.0483^2 m^2 / 4) = $-2.052*10^7$ Pa.

Now find the axial strain

$\sigma = E\ \epsilon$

$-2.052*10^7$ Pa = (35.0 *10^9 Pa) ϵ

$\epsilon = -5.863*10^{-4}$

Finally calculate the change in length

$\Delta L = \epsilon_L L_0$

ΔL = ($-5.863*10^{-4}$) (12.6 cm)= $-7.39*10^{-3}$ cm.

The negative sign indicates that the cylinder is shorter under the load. The change in length could have been found from $\Delta L=L - L_0$ by first finding the length L under the load and then subtracting the given value of L_0 .

EXAMPLE 1.8: A cylinder of rock with a Poisson' s ratio of 0.422 has a no-load length equal to 17.6 cm and no-load diameter of 6.20 cm. Under an axial compressive load, the axial strain is found to be -3.44 mm/m. Find the change in diameter under the load?

The transverse strain can be found from the axial strain and Poisson's ratio. However, first the axial strain must be put into dimensionless form as follows:

ϵ = -3.44 mm / m
ϵ = -3.44 $*10^{-3}$ m / m = -3.44$*10^{-3}$
$\boldsymbol{\epsilon_T = - \mu \epsilon_L}$
ϵ_T = - (0.422) (- 3.44$*10^{-3}$) = 1.452$*10^{-3}$
Now find the change in diameter :
$\boldsymbol{\Delta D = \epsilon_T D_0}$
ΔD = (1.452$*10^{-3}$)(6.20 cm) = 9.00$*10^{-3}$ cm = 90.0 μm
Note that 1 μm (micron) = 1$*10^{-6}$ m.

EXAMPLE 1.9: A limestone rock with Poisson's ratio of 0.276 and modulus of elasticity of 63.0 GPa. Find the values for other elastic moduli?

Shear Modulus or Modulus of Rigidity **(*G*)**
$G= E / 2(1 + \mu)$ = 63.0 GPa / 2(1+ 0.276) ; G = 24.7GPa
Bulk Modulus or incompressibility **(*k*)**
$k = E / 3 (1 – 2 \mu)$ = 63.0GPa / 3(1 - 0.276) ; k = 46.9GPa
Compressibility (K)
$K = 1 / k$ = 1 / 46.9GPa ; K = 0.0213 GPa^{-1}

1.3.2 Mechanical Properties of Rocks

Mechanical properties are also big described as engineering properties. These are the properties of rocks which help an engineer to fix the design parameters for a construction. Therefore, the most important property is the strength, which is the ability of a rock to resist an externally applied load. When a load is applied to the rock, the deformation should be within a limit and for this, deformability of the rock is also known. A knowledge of mechanical properties helps in analyzing the performance of the structure after the imposition of the load. Depending upon the type of loading and the stresses, the strength may be classified as:
1- Compressive strength 2- Tensile strength 3- Shear strength

1- Compressive strength

It is the stresses that are resulted from compressive forces causing contraction in the volume of rocks. There are two types:
a- **Unconfined (Uniaxial) Compressive Strength**: It is the most frequently used strength test for rocks in which a load on the rock acts in one direction only. There is no loading along an axis perpendicular to the loading axis. Rocks under compressive stresses fail in tension or shear depending on several factors such as moisture content and the associated swelling and many other factors. The compressive strength q_u is expressed as the ratio of peak load F, causing failure, to initial cross-sectional area A:

$$\sigma = q_u = F / A = N / m^2$$

b- Confined (Triaxial) Compressive Strength

Rocks in nature are seldom subjected to stresses in one direction , but subjected to stresses from three directions. Any point of rock mass under the earth surface is subjected to vertical stresses due to its overlying column load ($\sigma_1 = \gamma . Z$) in addition to the horizontal stresses. When a rock mass is subjected to an all-round pressure and if further subjected to additional vertical pressure, then strength exhibited by the rock mass is known as triaxial compressive strength. The lateral pressure acting on the rock is also known as hydrostatic (or lithostatic) pressure.

The usual procedure for conducting a triaxial compression test is applying vertical (σ_1) and horizontal stresses σ_3. At first, the specimen is subjected to confining pressure (horizontal stress) all round the rock cylinder (i.e. $\sigma_3 = \sigma_2$) which must be fixed, then applying the axial load (vertical stress) σ_1, until failure occurred, as the lateral pressure is held constant. In this test, at least, 3 cylindrical specimens of the same rock material are examined. This test also helps in the determination of shear strength by drawing Mohr's circles for the obtained values of σ_1 and σ_3 for each specimen. Mohr envelope gives the value of cohesion (***C***) and the angle of internal friction (Φ). The failure is occurred by the shear effect expressed by shear strength of rock mass using ***C*** and Φ parameters:

$$\tau = \sigma_n \tan \Phi + C \qquad (1.52)$$

Where τ is the shear strength and σ_n is the normal stress.

2- Tensile strength

It is the maximum tensile stress which a material is capable of developing. In nature, rock mass is rarely subjected to direct tension but it is subjected to tensile stresses. In the roofs and domes of underground openings, tension develops in the tensile zone of the rock mass. Tensile stresses also develops on the underside of a rock slab or a beam subjected to bending. Hence the knowledge of a tensile strength of the rock mass also necessary. Rocks are weak in tension. It has been found that rocks possess tensile strength which is about 10 % of its compressive strength. Since it is difficult to prepare rock samples for direct tensile tests, so tensile strength of rock samples are determined indirectly by other methods (***Brazilian and bending tests***). In general , the sedimentary rocks (such as limestones) are weak under tension stresses, so any constructed structure in or on these rocks must be reinforce by concrete.

3- Shear strength

It is the capacity of a rock mass to take a shear stress, or the maximum resistance to deformation due to shear displacement caused by a shear stress. Shear strength in a rock mass is derived from the surface frictional resistance along the sliding plane, interlocking between the individual rock grains and cohesion in the sliding surface of the rock. The pattern of joints, shear zones and faults in a rock mass reduces the effective shear strength of a rock mass. Hence, specially when the rock mass is supporting a concrete dam foundation, which is likely to experience a sliding force at the face due to an excessive water pressure, it is necessary to check the sliding resistance and shear resistance of the rock mass along the direction in which the maximum stress is expected to develop. The important problems in rock mechanics where a knowledge of shear strength of rock mass is needed, are the stability of rock slopes, stability of structures against sliding and stability problems of underground openings. Its test is similar to that of triaxial compressive strength which is indirect method for determining shear strength parameters C and Φ from Eq.(1.52):

$\tau = \sigma_n \tan \Phi + C$

This equation is true for clays, but for unconsolidated (loose) sediments such as sands , where $C = 0$, it becomes:

$$\tau = \sigma_n \tan \Phi \qquad (1.53)$$

The presence of water will reduce the normal stress due to pore water pressure, so the above equation becomes:

$$\tau = (\sigma_n - P_w) \tan \Phi + C \qquad (1.54)$$

where P_w is the pore water pressure and $(\sigma_n - P_w)$ is then called the effective normal stress σ_{eff} .

Other methods for shear tests are the direct shear tests which include shear box and cube shear tests.

EXAMPLE 1.10 : In a certain region and at a specific depth inside limestone rock, the results of triaxial compressive test are as follows:

Normal stress =13.49 MPa ; shear strength = 9.75 MPa ; cohesion = 1.17 MPa and internal friction angle = 40°. If a reservoir is suggested to construct at this site, what would be the pore water pressure and the effective normal stress?

$\tau = (\sigma_n - P_w) \tan \Phi + C$

9.75 MPa = (13.49 MPa – P_w) $\tan$ 40° + 1.17 MPa

9.75 MPa = (13.49 MPa – P_w) * 0.839 + 1.17 MPa

P_w = 3.26 MPa

$\sigma_{eff} = \sigma_n - P_w$

σ_{eff} = 13.49 MPa - 3.26 MPa

σ_{eff} = 10.23 MPa

1.3.3 Earth Stresses

Rocks under stress stores energy. This energy is the work done in deforming the rock. Any engineering project involving the removal or excavation of rock can result in the release of energy (rock bursts) should be avoided. Therefore, before undertaking construction projects requiring the disturbances of significant quantities of rock, the state of stress in the rock before work begins are called the ***initial stresses*** or the ***in-situ stresses*** which should be determined. These principal stresses are vertical and horizontal stresses. In shallow depths from earth surface, the horizontal stresses are greater than the vertical stresses. But at greater depths, the vertical and horizontal stresses are approximately equal from all direction and the rock mass will be under ***hydrostatic*** (***lithostatic***) pressure as shown in the below Figure (1.2).

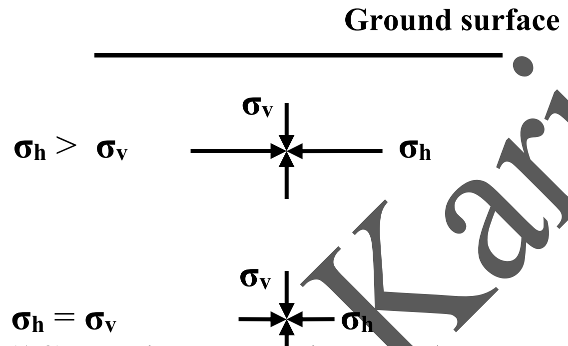


Fig. (1.2). Vertical and horizontal stresses beneath earth surface.

Although many factors bear on the state of stress at any point in a rock formation, the weight of the overlying materials (the overburden) is usually the most important one. The vertical stress can be calculated from the following relation:

$$\sigma_v = \rho\, g\, z \tag{1.55}$$

and in terms of the rock s unit weight γ:

$$\sigma_v = \gamma\, z$$

The horizontal stresses can be calculated from the lateral earth pressure (***k***) which represents the ratio of horizontal stresses to the vertical stresses.

$$k = \sigma_h / \sigma_v \tag{1.56}$$

and as $k = \mu / (1 - \mu)$ (1.57)

so $\mu /(1 - \mu) = \sigma_h / \sigma_v$ (1.58)

$$\sigma_h = \mu\, \sigma_v /(1 - \mu) = \gamma \Delta Z\, \mu / (1 - \mu) \tag{1.59}$$

Assuming Poisson's ratio μ for most rocks is between 0.2 – 0.3, thus the horizontal stresses would be in the range 1/3 – 1/4 of the vertical stresses.

Now suppose that beneath the earth surface, the rocks are arranged in several horizontal of different thicknesses and densities and that the stress is needed to be known at a depth beyond at least that of the first interface between two different layers. For example, suppose that the stress must be calculated at point ***P*** in Figure 1.3 below which will be as follows:

$$\sigma = \rho_1 g\, z_1 + \rho_2\, g\, z_2 + \rho_3\, g\, z_3 \tag{1.60}$$

where

$z_3 = z - (z_1 + z_2)$)

Note that z_3 is not the total thickness of layer 3. Thus the stress at any point is equal to the sum of stresses above this point:

$$\sigma = \Sigma\, \rho_i\, g\, z_i = \Sigma\, \gamma_i\, z_i$$

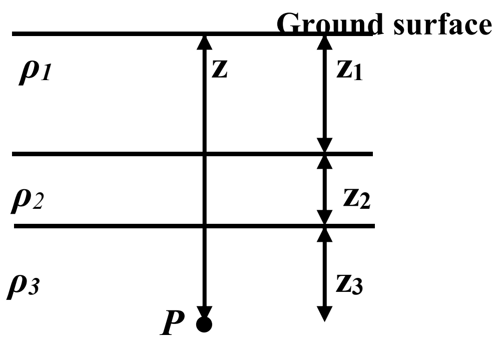


Fig. (1.3). Multilayered subsurface rock.

EXAMPLE 1.11: Find the vertical and horizontal stresses at depth 500 m from earth surface in sedimentary rocks with a density 2.75 gm/cm^3 and lateral earth pressure of 0.5.

$\sigma_v = \gamma z = 2.75 \text{ gm / cm}^3 * 500 \text{ m}$

$\sigma_v = (2.75/1000\text{kg}*9.8) \text{ N}*500 / 10^{-6} \text{ m}^3$

$\sigma_v = 13475000 \text{ Pa} = 13.475 \text{ MPa}$

$k = \sigma_h / \sigma_v$

$0.5 = \sigma_h / 13.5$

$\sigma_h = 6.75 \text{ MPa}$

EXAMPLE 1.12: Figure (1.4) shows a cross-section of rocks beneath the ground in a certain region. Three horizontal layers of rock (A, B and C) are shown together with their dry unit weights and depths from earth surface and also the depth of point P which lies in rock type C. It is required to find stress at point P?

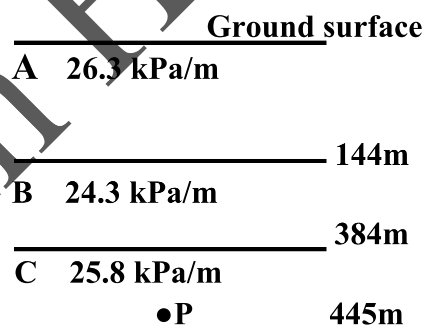


Fig. (1.4). Example 1.12.

$\sigma_p = \rho_A g z_A + \rho_B g z_B + \rho_C g z_C$

$\sigma_p = \gamma_A z_A + \gamma_B z_B + \gamma_C z_C$

$z_A = 144\text{m}$

$z_B = 384 \text{ m} - 144 \text{ m} = 240 \text{ m}$

$z_C = 445 \text{ m} - 384 \text{ m} = 61 \text{ m}$

$\sigma_p = (26.3 \text{ kPa/m})(144\text{m}) + (24.3\text{kPa/m})(240\text{m}) + (25.8\text{kPa/m})(61\text{m})$

$\sigma_p = 1.119 * 10^7 \text{ Pa}$

$\sigma_p = 11.2 \text{ MPa}$

REVIEW QUESTIONS AND PROBLEMS

1.1 Where do you make use of compressive strength values for rocks in civil engineering practice?

1.2 Is there any relation between the ultimate strength and the density of the rock?

1.3 If the compressive and tensile strengths of a rock are known, can you arrive at an estimation of the unconfined shear strength?

1.4 What is the range values of Poisson's ratio for rocks?

1.5 What is the effect of the degree of moisture saturation in the specimen on rock strength?

1.6 What is the relation between shear strength and θ?

1.7 What is the effect of of the size and shape of the aggregate particles on the bulk density and voids?

1.8 How do particles with different hardness values affect the strength of an aggregate?

1.9 A cylindrical sample of rock has a length of 37.7 cm and a diameter of 7.50 cm . The mass of the sample is 4747gm. Find the unit weight of the rock in kN/m^3. **(Ans. 27.9 kN/m^3)**

1.10 Find the unit weight of rock with a specific gravity of 3.08 .

1.11 Calculate the volume of a rock slab with unit weight 29.5 kN/ m^3 and mass 4570 kg. **(Ans. 1.52 m^3)**

1.12 A block of granite has edge lengths 1.24 m, 0.820 m, 0.933 m. It weighs 24.7 kN.
(a) Find the unit weight of the granite. (b) Find its specific gravity .

1.13 A block of rock has edge lengths 45.0 cm , 37.2 cm , 12.8 cm. Its porosity 38.4%. Find the total volume of the pores in the block. **(Ans. 8230 cm^3)**

1.14 The specific gravity of a rock is 2.94 with a porosity of 0.344. Calculate the specific gravity of the grains.

1.15 A 12.74 m^3 block of rock has a porosity of 26.40%. What is the volume of this rock after the block is crushed just sufficiently to close all the pores?

(Ans. 9.38 m^3)

1.16 A block of rock has edge lengths 1.22 m, 2.40 m, 1.81 m. When dry its mass is 14.7 mg; when saturated with water its mass is 16.6 mg. Find the porosity of the rock .

1.17 A slab of rock has a volume of 5.56 m^3 and a porosity of 0.417 . It is saturated with oil of density 0.620 gm/cm^3. Find the weight of the oil in the slab.

(Ans. 14.1 kN)

1.18 A rock saturated with oil has a unit weight of 29.3 kN/m^3. When dry the rock has a unit weight of 26.4 kN/m^3. The porosity of the rock is 0.370. Determine the density of the oil.

1.19 A cube of chalk with porosity 38.4% has an edge length of 1.40 m. The chalk is crushed, closing all the pores, and then reshaped into a cube. What is the edge length of the new cube? **(Ans. 1.19 m)**

1.20 Calculate the porosity of a rock that is 50% quartz , 50% muscovite, and that has a bulk density of 2.0 gm/cm^3. **(Ans. 0.273)**

1.21 Calculate the porosity of a rock that is 50% quartz, 50% muscovite , and that has a bulk density of 2.0 gm/cm^3 when saturated with water.

1.22 A cylindrical sample of rock has a diameter of 8.48 cm and a length of 14.6 cm. When dry it weighs 22.8 N.
(a) Find the porosity of the rock.
(b) What is the volume of the water in the sample when saturated with water? **(Ans. a- 64.4% ; b- 531 cm^3)**

1.23 A block of oil shale with a volume 0.774 m^3 is saturated with 0.311m^3 of oil. The unit weight of the saturated oil shale is 27.8 kN/m^3. After all of the oil has been driven out of the rock, the unit weight of the rock is 25.20 kN/m^3. Find the specific gravity of the oil. **(Ans. 0.660)**

1.24 A force of 17.200 N is exerted perpendicularly against a surface of area 0.136 m^2. Find the stress. **(Ans. 126 kPa)**

1.25 A cylinder of rock has a length of 5.82 cm and a diameter of 2.14 cm. Oppositely-directed axial forces of 14.8 kN each are exerted against the ends. Find the stress. **(Ans. 41.1 MPa)**

1.26 In a certain region where subsurface rocks have density 3.08 gm/ cm^3 , what is the lithostatic stress at a depth of 4.75 km.

(Ans. 143 MPa)

1.27 By how much does the vertical stress at a depth of 1.22 km exceed the stress at a depth of 840 m if the underground rocks have density 2420 kg/ m^3? **(Ans. 9.01 MPa)**

1.28 At what depth below the ground surface in Fig. (1.5) is the value of the vertical stress equal to 6.08 MPa? **(Ans. 235 m)**

Fig . (1.5). Problem 8.28.

1.29 Figure (1.6) shows three subsurface layers of rock ; the thicknesses of the layers are given, together with the densities of the two upper layers. If the stress at a depth of 1.2 km is 34.4 MPa.
(a) Find the stress at a depth of 350 m.
(b) Find the stress at a depth of 600 m.

(Ans. a- 8.23 MPa ; b- 14.8 MPa)

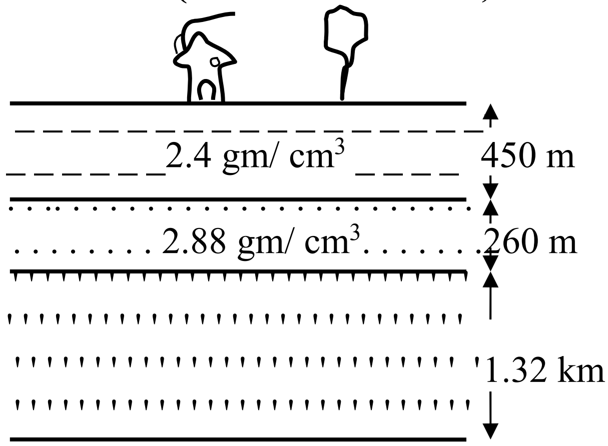


Fig. (1.6) . Problem 1.29.

2. Surface Water and Rivers Geologic Work

2.1 Introduction

The water is found mainly in the oceans, occurring 70.8 % of the earth surface which are the main source of water cycle in nature. Water that falls as ***precipitation*** entered the atmosphere by ***evaporation*** mainly from the oceans. When the precipitation is over an ocean area, the water returns directly to its source. Sometimes, water can be evaporating back into the atmosphere . The water that does not evaporate either infiltrates into the ground (***infiltration***) or forms surface water ***runoff*** (Fig. 2.1). The water that does not infiltrate into the ground , the ground water runoff, may be absorbed by the roots of plants which vapor back into atmosphere , this process is called ***transpiration***. Both processes are called together ***evapotranspiration*** .

Thus:

Amount of falling water=Evaporated water+Infiltrated Water+Runoff Water

or (2.1)

Total precipitation = Evaporation + Groundwater increment (by infiltration) + Direct runoff

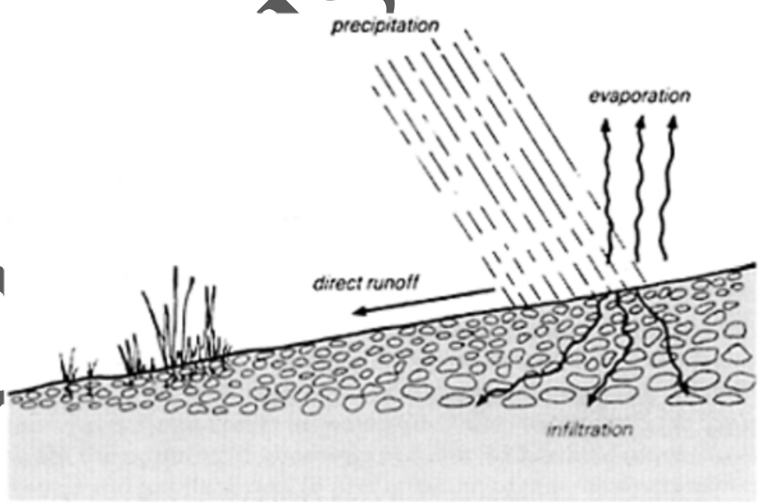


Fig. (2.1). A globule of water precipitated as rain is dispersed by evaporation, by direct runoff and by infiltration into the ground. The division between runoff and infiltration is determined by the relative resistance to flow along either path. Vegetation and steepness of slope influence the flow into streams, and the near-surface permeability controls infiltration.

9.2 Water Movement in Rivers

Water usually moves toward the gradient. There are several factors affect this movements, of these are: velocity (***V***), gradient (***S***) and channel (river) dimensions. Besides, there are other factors reduce water velocity such as the friction between water and the bottom (bed) of the river and the friction among water particles themselves and with the suspended materials.

The ***driving force*** of a river equals water mass (***M***) times the gradient (***S***),where the water mass ***M*** is the product of the multiplication of the river cross-sectional area (***A***) , length of the river (***L***) and water density (***ρ***). Thus:

$$M = AL\rho \qquad (9.2)$$

$$\text{The driving force} = AL\rho S \qquad (9.3)$$

The other force which acts as the opposite force to the driving force is the ***total friction force*** which is equal to the friction per unit area (F_A) times the area of the river bottom. The later is equal wetted perimeter (***P***) times the length of the river (Fig. 2.2), $F_A\ PL$. The two forces become equal when the flow is at a constant velocity then:

$$AL\rho S = F_A\ PL \qquad (9.4)$$

Thus the friction per unit area is:

$$F_A = \rho S\ (A/P) \qquad (9.5)$$

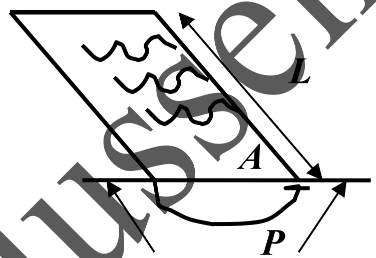


Fig. (2.2). A cross-section for a river.

The main variations causing variations in river velocity are gradient (***S***), shape of the river cross-section (***A /P***), and degree of roughness of the surface on which water flows which is expressed by ***roughness coefficient*** (***n***). Thus the general relation is:

$$V^2 \propto \frac{S}{n}\,\frac{A}{P} \qquad (9.6)$$

Where the hydraulic radius ***R = A / P***.

It is noticed from the above relation that the geometric shape of the river (***A / P***) affects the velocity, as the wetted perimeter decreases the velocity increases for decreasing friction. Besides, as roughness of the surface decreases the velocity increases too as shown in Figure (2.3).

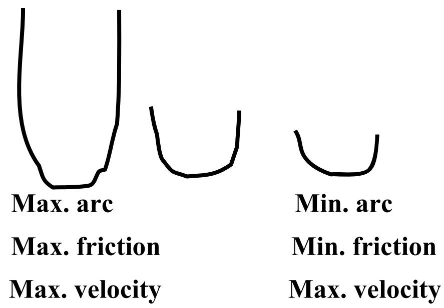


Fig. (2.3). Influence of channel shape.

2.3 Discharge

It is the amount of water flowing past a certain point in a given unit of time. This is usually measured in cubic meter per second by multiplying cross-sectional area (***A***) by its velocity (***V***).

$$Q = A.V = (wd).V \tag{2.7}$$

Where ***Q*** is the discharge of the river; ***w*** is the average width of the river and; ***d*** the average depth of the river.

Two types of discharges are presented in rivers, these are:

1- Laminar flow: It characterizes by its sheet flow without turbulence which occurred in smooth and straight channels and it is very slow.

2- Turbulent flow: It is usually occurred in rivers and characterizes by its circular movement.

2.3.1 Relation between River Discharge and other Hydraulic Parameters

Rivers, after a specific time of their formation, are being in a ***graded system***. A ***graded river*** has the correct slope and channel characteristics necessary to maintain just the velocity required to transport the material supplied to it. On the average, graded system is not eroding or depositing material but is simply transporting it. Once a river has reached this state of equilibrium, it becomes a self-regulating system in which a change in one characteristic causes an adjustment in the others to counteract the effect. The graded river characteristic is that any change in any of the controlling factors will cause a displacement of the equilibrium in a direction that will tend to absorb the effect of the change.

The civil engineer, however, is under compulsion to produce quantitative answers in numerical terms. If this compulsion makes it necessary to simplify the problem, it is important that the engineer's judgment be guided by the best possible comprehensive of even the physical factors which are not fully expressed in his computations. The first simplification which the engineer hastens to adopt is an assumption that the erodible banks and bed of the river have been transformed into a rigid conduit. According to this assumption, all the hydraulic parameters become immediately available for establishing desirable values of velocity, depth, or width. Only discharge remains an independent variable. Discharge is equal to the sum of the runoff from the watershed plus the contributions from groundwater flow. Besides, water flows over soil much more rapidly than it flows through soil. Consequently, flood flow is primarily the result of runoff, while the discharge of a river is maintained by seepage from underground reservoirs.

Thus the discharge of a river at any fixed location is usually not constant with time. The discharge can vary significantly, sometimes over a time of only a few hours. For example, if there has been heavy precipitation, or melting of snow, upstream, it is to be expected that much of the runoff generated will, after some delay, show up as increased discharge of the river, it can be quite important to know the river will respond, through changes in the values of depth, width, and current velocity. Thus, these changes cannot be predicted on the basis of the above mentioned simple equation alone, for one equation cannot generally be solved for three unknowns, so more equations are needed.

2.3.2 Determination of Discharge

As mentioned above, the discharge represents the relation between river cross-sectional area and the velocity:

$Q = A.V = (wd).V$

As mentioned above ***w*** and ***d*** should be the average width and depth of the river as rivers seldom run in straight line paths, but meander, and thus the velocity ***V*** of the water has different values within a cross section. It is also assumed that the river flow is not turbulent.

Of the first relations related between river velocity and its gradient and hydraulic gradient is ***Chezy formula***:

$V = C\,(R.S)^{1/2}$; (2.8)

$Q = C\,(R.S)^{1/2}\,A$ (2.9)

Where ***V*** is the mean velocity ; ***R*** is hydraulic mean radius ; ***C*** is the roughness coefficient which varies with the characteristics of the channel , and ***S*** is the slope (gradient) of the channel.

There are many methods for determining the factor ***C***, of which is ***Manning's formula*** which is an exponential relation between R and ***C*** and has a wide applications:

$C = (1/n)\,R^{1/6}$ in Metric system (2.10)

$C = (1.49/n)\,R^{1/6}$ in English system (2.11)

Where ***n*** is the roughness or the channel surface representing the dissipating energy of the river.

Combining Chezy and Manning equations, we get the following equations:

Metric System		English System	
$Q = (1/n)\,A\,R^{2/3}\,S^{1/2}$		$Q = (1.49/n)\,A\,R^{2/3}\,S^{1/2}$	(2.12)
$V = (1/n)\,R^{2/3}\,S^{1/2}$		$V = (1.49/n)\,R^{2/3}\,S^{1/2}$	(2.13)
$S = (n^2.\,V^2)\,/\,R^{4/3}$		$S = (n^2.\,V^2)\,/\,2.22\,R^{4/3}$	(2.14)

EXAMPLE 2.1: A stream starts out 2000m a.s.l. and travels 250 km to the sea. What is its average gradient in m / km? Suppose that the stream developed extensive meanders so that its course lengthened by further 250 km. Calculate its new gradient. How does meandering affect gradient?

$H_1 = h / S =$ 2000 m / 250 km = 8 m / km

$S = S_1 + S_2$ = 250 km + 250 km = 500 km

$H_2 = h / S =$ 2000 m / 500 km = 4 m / km

It is obvious that river meandering reduces river gradient and also its velocity, in our example the gradient reduces to a half.

EXAMPLE 2.2: Three water channels with different shapes, the first is semicircular cross-sectional area with a radius of 11.25 m, the second is shallow rectangular with a depth of 2 m and width 100 m while the third which is deep rectangular with depth 5 m and width 40 m. Find:

1- Wetted Perimeter in each channel.
2- Hydraulic radius for each channel.
3- Discuss your results .

$\boldsymbol{A_1 = \pi r^2} = \pi * (11.25 \text{ m})^2 = 397.4 \text{ m}^2$

Semi circle = A_1 $/ 2 = 397.4 \text{ m}^2 /2 = 198.7 \text{ m}^2 = 200 \text{ m}^2$

$\boldsymbol{P_1 = 2\pi r / 2} = 2 * 3.13 * 11.25 \text{ m} / 2 = 35.32 \text{ m}$

$\boldsymbol{R_1} = \frac{A_1 /2}{P_1} = 198.7 \text{ m}^2 / 35.32 \text{ m} = 5.62 \text{ m}$

$\boldsymbol{A_2 = W_2 * d_2} = 100 \text{ m} * 2 \text{ m} = 200 \text{ m}^2$

$\boldsymbol{P_2} = 100 \text{ m} + 2 \text{ m} + 2 \text{ m} = 104 \text{ m}$

$\boldsymbol{R_2 = A_2 / P_2} = 200 \text{ m}^2 / 104 \text{ m} = 1.923 \text{ m}$

$\boldsymbol{A_3 = W_3 * d_3} = 40 \text{ m} * 5 \text{ m} = 200 \text{ m}^2$

$\boldsymbol{P_3} = 40 \text{ m} + 5 \text{ m} + 5 \text{ m} = 50 \text{ m}$

$\boldsymbol{R_3 = A_3 / P_3} = 200 \text{ m}^2 / 50 \text{ m} = 4.0 \text{ m}$

Discussion: The cross-sectional shape of a channel determines the amount of water in contact with the channel and hence affects the frictional drag. The most efficient channel is one with the least perimeter for its cross-sectional area. From above results, the three types of channels have equal cross-sectional area. But the first channel (with a semicircular shape) has less wetted perimeter and higher hydraulic radius and hence less frictional drag, as a result the water will flow with higher velocity and discharge. While, the second channel (shallow rectangular shape) has longer perimeter, lower hydraulic radius and hence more frictional drag , as a result the water will flow with lower velocity and discharge. Whereas, the third channel (deep rectangular shape) will be between the first and second channels**.**

EXAMPLE 2.3: Circle river flows in a channel with a semicircular cross-section of width 34.2 m . After a storm upstream , the river is found to be just within its bank and flowing at 13.5 m / s: see Fig (2.4).

a- calculate the discharge.

b- Find the volume of water that flows through a cross-section in 2.0 hrs.

Answer

a- With the cross-section being a semicircle, we must use one- half the area of a circle. Since the width of the river is the diameter ***D*** of the semicircle. So the discharge of the river:

$\boldsymbol{Q = 1 /2 [(1 / 4) \pi D^2)] V}$

$\boldsymbol{Q} = (1 / 8) \pi (34.2 \text{ m})^2 (13.5 \text{ m} / \text{s})$

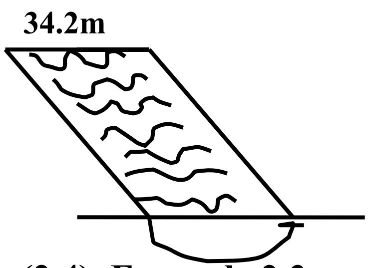


Fig. (2.4). Example 2.3

Q = 6200 m^3 / s

b- With ***Q*** evaluated, the volume of water that passes can be found, but ***Q*** is in m^3/s and ***T*** =2.0 hrs, so the time units must be reconciled. Since 1***hr***=3600***s***:

V = Q T

V = (6200 m^3 / *s*) [(2)(3600s)]

V = 4.46 * 10^7 m^3

EXAMPLE 2.4: A river with a width 120 m, depth 7 m, discharge 9065 m^3 / s, gradient 0.002 and roughness coefficient 0.018. How long is the wetted perimeter, hydraulic radius and the river velocity?

$Q = (1/n)\, A\, R^{2/3} S^{1/2}$; Where $R^{2/3} = A^{2/3} / P^{2/3}$

9065 m^3 / s = (1/ 0.00185) * 840 m^2 *[(120m * 7m)$^{2/3}$ / $P^{2/3}$]* (0.002)$^{1/2}$

9065 = (1/ 0.0018) * 840* (89.02 / $P^{2/3}$) *0.044

$P^{2/3}$ = 3290.18 / 163.17 = 20.164

P= 90.545 m

R =A / P = (120m * 7m) / 90.545 = 9.277 m

Q= V . A

9065 m^3 / s =***V*** * 840 m^2

V= 10.79 m / s

EXAMPLE 2.5: A stream with rectangular cross-sectional area, its gradient 0.0036. Find its velocity at the cross-section where the width 28.4 m and the depth 9.3 m, knowing that Chezy coefficient (roughness coefficient) is 33 √m /s.

First find the hydraulic radius (***R***) and as the cross-section shape is rectangle, so the wetted perimeter is (twice depth + width) :

R= A / P = (28.4 m)(9.3 m) / 2 (9.3m) + 28.4 m

R = 5.62 m

Using Chezy Formula to find water velocity;

$V = C\,(R.S)^{1/2}$

V = (33 √m /s) [(5.62m)(0.0036)]$^{1/2}$

V = 4.7 m /s

2.4 Mechanism of Rivers Geologic Work

The work of a river includes three main processes, erosion, transportation and deposition. These activities go on simultaneously in all river channels, even though they are presented individually here.

1- Erosion

Rivers and streams are eroding rocks and sediments in their own channels (from both bottom and sides). If a channel is composed of bedrock , most of the erosion is accomplished by ***abrasion*** action of water with sediments which is the main process. While, in channels consisting of unconsolidated materials, considerable lifting can be accomplished by the ***impact*** of water alone. Other processes of erosion are ***solution***, ***corrosion*** and ***slope wash.***

It is found that the coarse-grained fragments such as gravels are moved by direct impact. The maximum diameter (D_{max}) of such grains could be moved is directly proportional with the velocity square (V^2) by the relation:

$$D_{max} = K\, V^2 \qquad (2.15)$$

Where K is the constant of proportionality.

It is obvious that the river velocity is affected by the main factors gradient; shape, size, and roughness of the channel; and discharge.

2- Transportation

The ability of a river to carry its load is established using two main criteria. First, river ***capacity*** which is the maximum load a river can carry. The greater the amount of water flowing in the stream, the greater the river capacity for hauling sediments. Second, ***river competence*** which is a measure of the maximum size of particles it is capable of transporting. The river velocity determines its ***competence***. If the velocity of a river doubles, the impact force of water increases four times: if the velocity triples, the force increases nine times; and so forth. Hence, the huge boulders which are visible during low-water stage and seem immovable can, in fact, be transported during flood stage because the river increased velocity. Every river has a specific energy which mainly depend upon its velocity and size . and as the size of the river is constant compared with its velocity which is varied from season to another, in addition to other factors such as gradient and roughness of the channel. Figure (2.5) shows the general change in river characteristics from head to mouth.

Rivers and streams transport their load of sediments in three ways:
a- in solution (dissolved load) ; b- in suspension (suspended load) and c- along the bottom (bed load).

a- Dissolved load

The dissolved load is brought to the rivers by groundwater and to lesser degree is acquired directly from soluble rock along the river course. The quantity of material carried in solution is highly variable and depends upon such factors as climate and geologic setting. The most important soluble materials are calcium (or sodium and magnesium) carbonates, chlorides and sulphates.

b- Suspended load

Most rivers, but not all , carry the bulk of their load as suspended load. Usually only fine sand-, silt-, and clay-sized particles can be carried this way, but during flood stage larger particles are carried as well. Also during flood stage, the quantity of material carried in suspension increases dramatically.

c- Bed load

A portion of a river load of solid material consists of sediments that is too large to be carried in suspension. These coarser particles move along the bottom of the river and constitute the bed load. The particles composing the bed load move along the bottom by rolling, sliding and saltaion. Sediment moving by saltation appears to jump or slip along the river bed. This occurs as particles are propelled upward by collisions or sucked upward by the current and then carried downstream a short distance until gravity pulls them back to the bed of the river. Particles that are too large or heavy to move by saltation either roll or slide along the bottom, depending upon their shape.

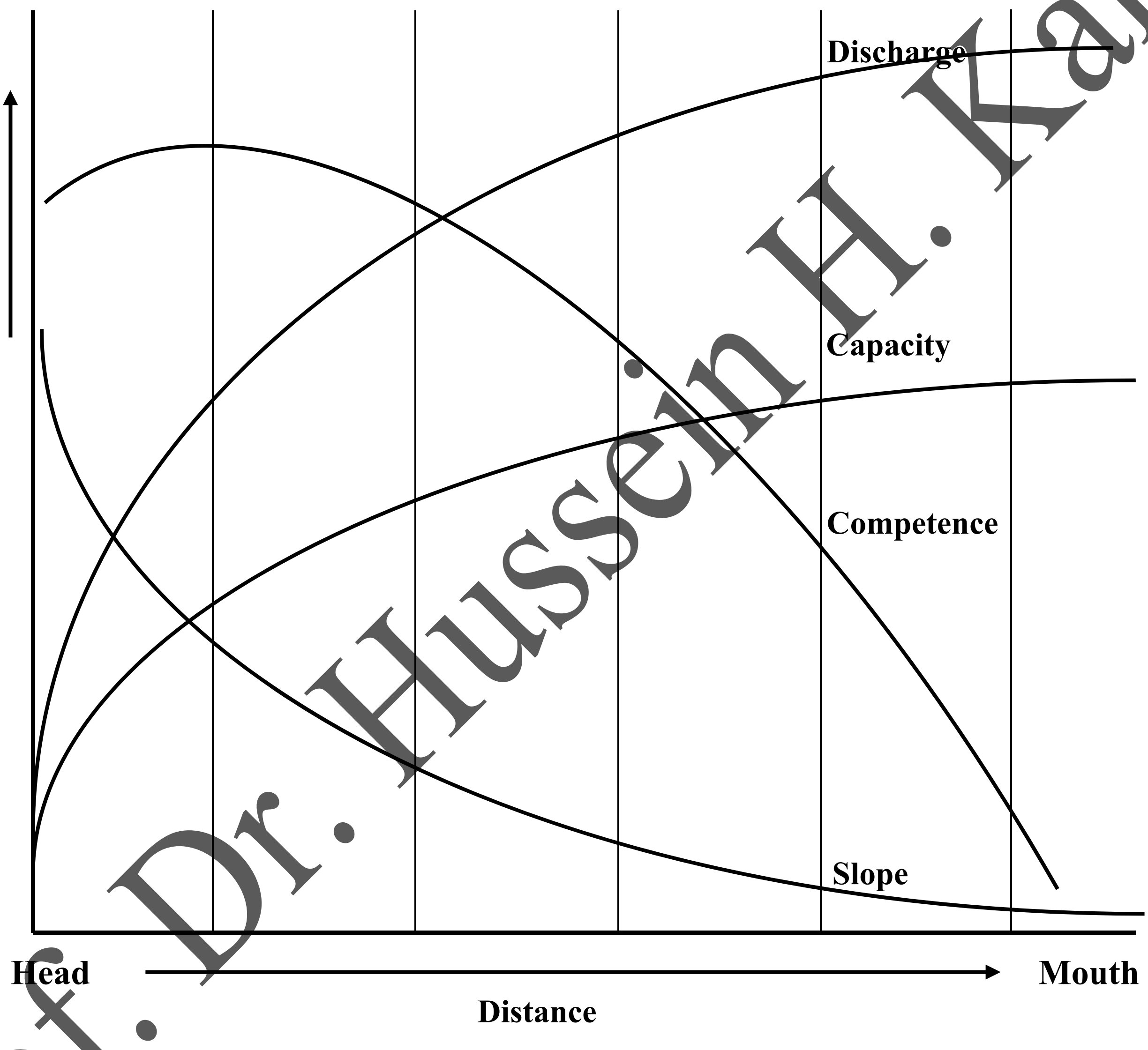


Fig.(2,5). The general change in river characteristic from head to mouth.

3- Deposition

Whenever a river velocity is reduced , its competence is also lowered. Consequently, some of the suspended particles begin to settle out. Sediment deposited in this manner is called ***alluvium*** (or ***alluvial deposits***). In some cases, when the river velocity is reduced there will be a ***sorting*** of the sediments deposited. With the reduction in the river velocity, the coarse particles such as gravels will be deposited and the river will transport the finer sizes such as sands, silts and clays. Finer sizes remain in suspension and are carried along by the current. With further reduction of velocity will result in further deposition of the coarsest particles remaining , and so on. Thus, the river will deposit its coarsest sediments first, and carry the finer ones along to be dropped , in succession, as the velocity decreases in the lake or sea. We find, as a result, that the sediments are sorted according to size, the coarsest may be deposited very far from the finest. Then, the marine sediments will consist of sands, silts and clays with the absence of gravels except rare cases. But the river sediments will consists of gravels, sands, silts and clays.

The main types of river deposits are:

1- Alluvial fans: When mountain rivers reach a plain, their gradient is abruptly lowered and they immediately dump much of their load. Usually the coarse material is dropped near the base of the slope, while the finer material is carried farther out on the plain.

2- River levees: Rivers that occupy valleys with broad, flat valley floors on occasion build a landform called a natural river levee that parallels its channel. Natural levees are built by successive floods over a period of many years. These levees consist of mainly fine sand, silts and some clays.

3- Flood plains: The plains that are found near the estuaries and along side the river valley resulted from river deposits during successive floods. They consist mainly fine sand, silts and clays.

4- Channel deposits: The deposits that are formed in the channel of the river are called ***river islands*** or ***sand bars***. Their formations are due to the reduction in the river velocity and its gradients or to the presence of some natural obstructions in the river channel. Thus, they are found as a straight lines between river meanders. Other cause of their formation is the succession between the continuous cut and fill processes. These landforms consist of varied grain-sized particles.

5- Meander deposits: It is the main characteristics of the flood pains which are well known in the central and southern parts of Tigris and Euphrates rivers in Iraq, especially at Al-Ziwiya and Al-Jadiriya regions in Baghdad. Rivers that flow upon floodplains and move in sweeping beds are called ***meanders.*** Their causes are related to nature and characteristics of the river. Meanders have sinuous shape similar to the mathematical sinuous curve. Meanders continually change position by eroding sideway and slightly downstream. The sideways movement occurs because the maximum velocity of the river shifts toward the outside of the bend, causing erosion of the outer bank (Fig.2.6). At the same time, the reduced current at the inside of the meander results in the deposition of coarse sediments, especially sands. Thus by eroding its outer bank and depositing materials along its inner bank, a river moves sideways without changing its channel size. Due to the slope of the channel, erosion is more effective on the downstream side of a meander. Therefore, in addition to growing laterally, the bends also gradually migrate down the valley.

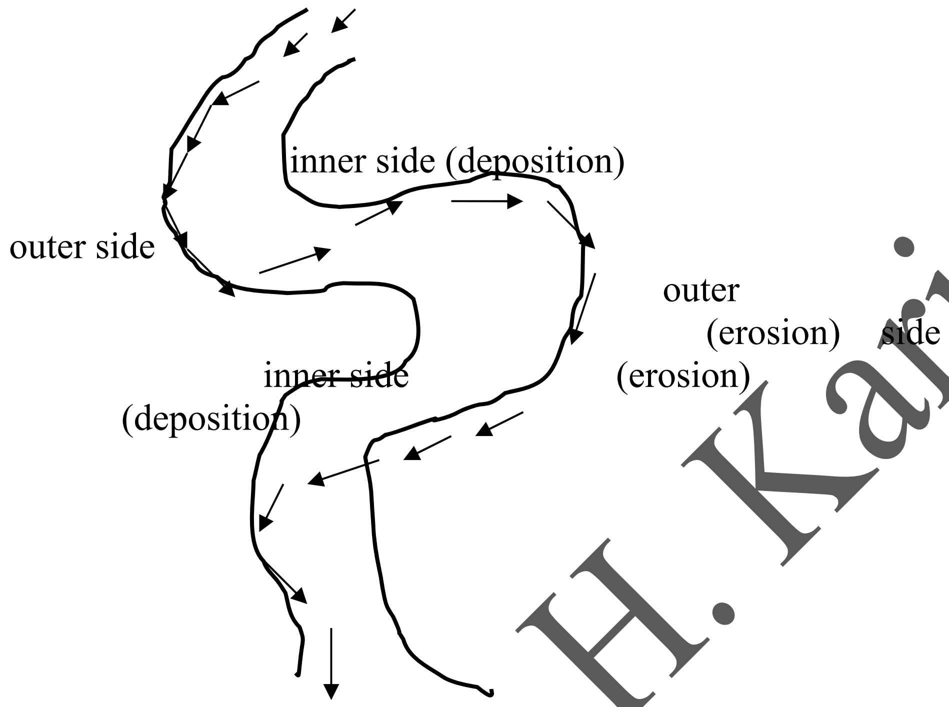


Fig. (2.6). River meanders.

Sometimes the downstream migration of a meander is slowed when it reaches a more resistant portion of the floodplain. This allows the meander upstream to ***"catch up"***. Gradually the ***neck*** of land between the meanders is narrowed. When they get close enough, the river may erode through the narrow neck of land to the next loop. The new, shorter channel segment is called a ***cutoff*** and, because of its shape , the abandoned bend is called an ***oxbow lake*** (Fig.2.7). When the water of the oxbow is dried the a ***meander scar*** occurred.

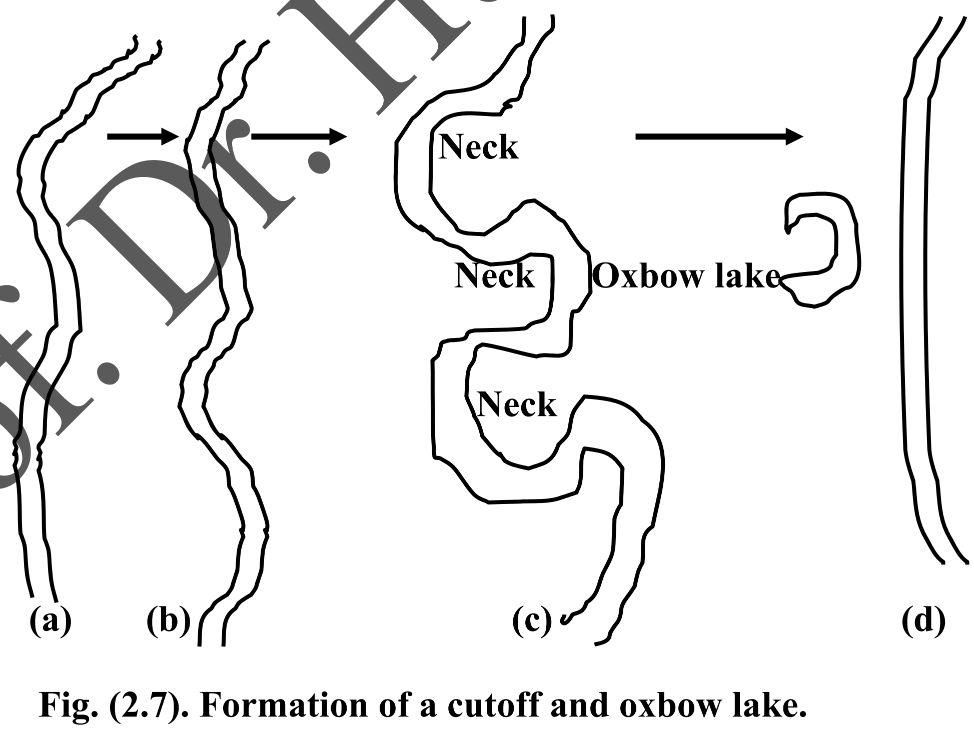


Fig. (2.7). Formation of a cutoff and oxbow lake.

6- Delta Deposits: When a river enters the relatively still waters of an ocean or lake , its forward motion is quickly lost, and the resulting deposits form a ***delta*** which grows into the idealized triangular shape of the Greek letter (**Δ**), for which it was named. The finer silts and clays will settle out some distance from the mouth into nearly horizontal layers called ***bottomset beds***. Prior to the accumulation of bottomset beds, ***forest beds*** begin to form. These beds are composed of coarse sediments, which is dropped almost immediately upon entering a lake or ocean, forming sloping layers. The forest beds are usually covered by thin, horizontal ***topset beds*** deposited during flood stage. As the delta grows outward , the gradient of the river is continually lowering. Some conditions must be presented to form the delta ; the river must transport large loads of sediments, high currents must be absent in the sea or lake (i.e. still water) and low gradient of the shore line (Fig. 2.8).

7- River Terraces: They are elevated flat surfaces occurred along one side of some river valleys. A new flood plain may be built with lower elevation with respect to the first one. This process is repeated many times and the result is a number of river terraces where the older terrace is the higher. This landform is common in northern Iraq and along the valleys of the Upper and Lower Zab, Adhim and Tigris rivers which could be a result of the earth movement that causes the continuous uplift (Fig. 2.8).

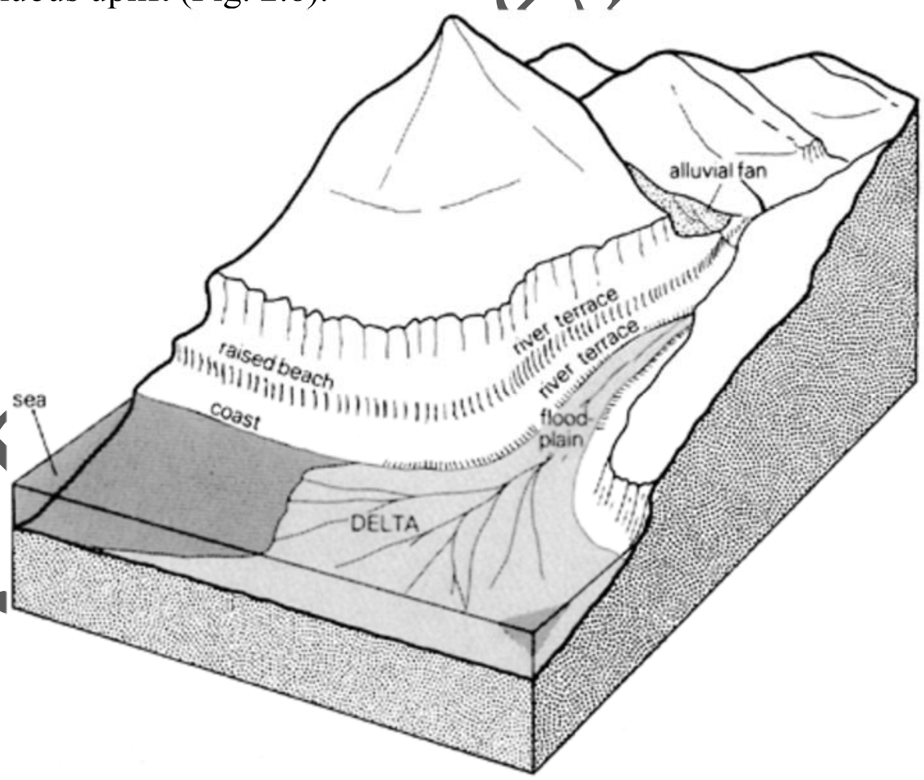


Fig. (2.8). The mountain stream has deposited the coarser part of its load as an alluvial fan where the gradient changes suddenly at the plain . Deposition of finer sediment has taken place at the coast to form a delta. As the river winds towards the coast, it cuts a broad floodplain. The river terrace is an older floodplain, which is now well above the river bed.

2.5 Stages of River Development

A- Youth Stage: This stage is usually near the river head and characterizes by its high water velocity and carrying all the eroded materials during its course. The destruction activities are larger than construction ones. The main characteristics of this stage are:
1- narrow V-shaped valleys ; 2- the primary work of the river is downcutting rather than from its sides ; 3- the presence of rapids and waterfalls ; 4- the absence of food plains ; 5- the absence of large meanders .

B- Maturity Stage: In this stage, the river widen its valley due to the continuous erosion with lower gradient due to this wideness. There is an equilibrium between erosion and deposition. It characterizes by the followings: 1- wide V-shaped valleys with flat floors ; 2- they have been widened by lateral (side- to- side) erosion ; 3- the absence of rapids and waterfalls ; 4- they are ready for formation of flood plain ; 5- formation of meanders and oxbow lakes.

C- Old Stage: This stage is usually occurred near estuaries with decreasing erosion and increasing deposition. The river valley is wide with low gradient and the main channel dividing into several smaller ones. This stage is characterized by the followings:
1- very wide valleys with flat floors ; 2- the presence of food plains and river deposits **;** 3- the complete absence of rapids and waterfalls **;** 4- decrease in gradients and erosion with an increase in deposition **;** 5- the appearance of several smaller branches (distributaries) and formation of delta.

REVIEW QUESTIONS AND PROBLEMS

2.1 Describe the movement of water through the hydrologic cycle . Is there more than one path which precipitation may take after it has fallen?

2.2 Why does the downstream portion of a river have a gentle gradient when compared to the headwater region?

2.3 Define stream load and What factors control it? In what way does a stream transport its load?

2.4 Differentiate between competency and capacity?

2.5 In what way is a delta similar to an alluvial fan? In what way are they different?

2.6 Why must the height of many artificial levees be increased periodically?

2.7 What is the purpose of artificial cutoffs?

2.8 Why is it possible for a youthful valley to be older (in years) than a mature valley?

2.9 Do streams flowing in mature and old - age valleys make good political boundaries? Explain.

2.10 A stream starts out 2000 meters above sea level and travels 250 kilometers to the ocean. What is its average gradient in meters per kilometer?

2.11 Suppose that the stream mentioned in Question (9.10) developed extensive meanders so that its course was lengthened to 500 kilometers. Calculate its new gradient . How does meandering affect gradient?

2.12 When the discharge of a stream increases , what happens to the stream velocity?

2.13 A stream has a width of 5.20 m, depth 3.70 m, and current speed 63 cm/s.
- a- Calculate the discharge.
- b- How much water flows through a cross section in 1.0 hr. ? Give the volume and weight of the water.

2.14 Water is flowing in a rectangular channel of 39.4 m wide and inclined at $0.180°$. The Chezy coefficient equals $51.2 \sqrt{m}$ /s. The water is flowing at 6.27 m/s. Find the depth of the water in the channel.

(Ans. 6.30m)

2.15 A river has a cross section closely resembling a trapezoid. The bed width is 17.7 m, and the two sides slope upwards and outwards from the bed, each at an angle of $32.0°$ to the horizontal. The Chezy coefficient equals $35.0\sqrt{m}$ /s and the river's inclination is $0.650°$. Find water's speed when the depth of the water in the middle of the stream is 5.22 m.

(Ans. 7.11 m)

3. Subsurface (Ground) Water

3.1 Sources of Groundwater

1- Meteoric water: Most of this water is and has its source in precipitation of rain, snow and dew. The last of these is important to many desert areas where survival of an animal, or of a human community, may depend on the ability to trap and use dew.
2- Magmatic (Juvenile) water: It may be a product of recent volcanic activity and have separated from magma. This juvenile water is often heavily contaminated with dissolved minerals and gases.
3- connate water: It may have been isolated from the cycle for millions of years as connate water, sealed in porous sediment by surrounding impermeable rocks, or lying near-stagnant deep within a groundwater basin.

Juvenile and connate waters form only a minor part of the water present in the uppermost kilometer of the solid earth.

3.2 Classification of Rocks with Respect to Groundwater Studies

The behavior of groundwater as it circulates through layers of different permeability underground and is eventually discharged on the surface or at the coast produces a range of phenomena. For simplicity in describing these, rocks are grouped into three categories.
1- **Porous and permeable rocks**: Those that have relatively high permeabilities because the pores of the rock are large. A body of ***pervious*** (and permeable) rocks capable of yielding groundwater is called an ***aquifer,*** such as sands.
2- Non Porous and pervious: Those that have relatively high permeabilities because well jointed, are referred to for ***brevity*** as ***pervious***, such as limestones. (This term is conventionally used to describe high fluid conductivity of a rock in which the voids are mainly secondary, that is, where there is appreciable flow along joints, fissures and bedding planes, in contrast to ***"permeable"***, which is then restricted to a description of primary ***pore permeability***).
3- Porous and Impermeable rocks: Rocks with porous and low permeability and no fissures are referred to in the following account as impervious, they form ***aquicludes***, such as clays.
4- Non porous and non- pervious: Such as quartzites and porcellanceous limestones.

3.3 Porosity, Hydraulic Conductivity and Permeability of Rocks

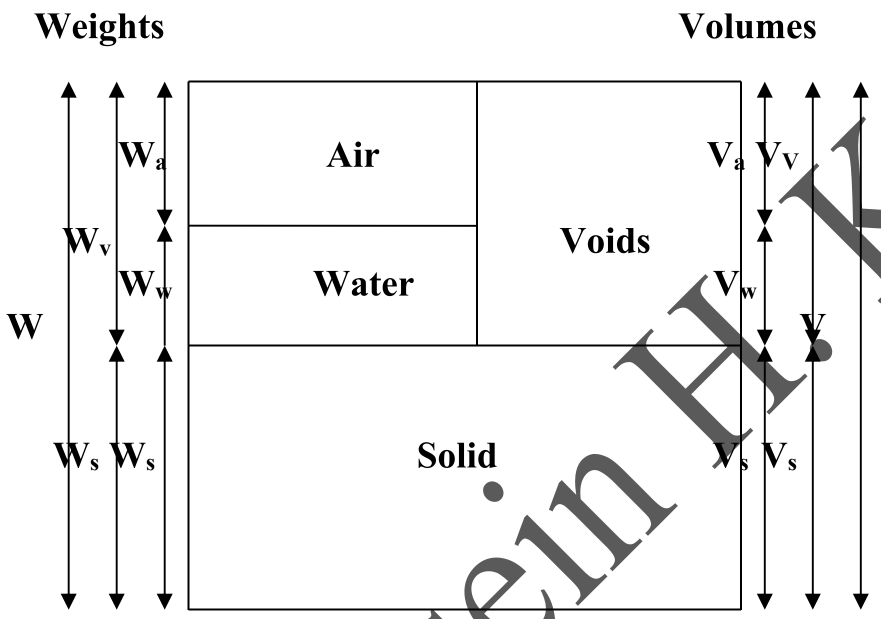
The relative ease of flow of water through a soil or rock is ***hydraulic conductivity (K)*** (referred to as the ***coefficient of permeability*** in some older texts). When water, oil or gas is flowing through the voids in a rock, its hydraulic conductivity is dependent on its viscosity and on other factors as well as on the properties of the rock. The properties of the rock ***alone*** that affect ease of flow are defined by its ***intrinsic permeability (k)***, (usually shortened to ***permeability***). This index property is used for definition in the petroleum industry, where more than one fluid is of interest, but is seldom used in hydrogeology. Rocks must have voids (that is, they must be porous), in order to have permeability and hydraulic conductivity, but the relationship between porosity and permeability depends mainly on the size of the voids rather than on their frequency. Clay has high porosity combined with low permeability. In an idealised terrigenous rock or soil in which the grains were spherical, of one size, and uncemented, the permeability would increase with the ***(diameter)***^{2} of the grain size. In real terrigenous soils there is a corresponding exponential increase in permeability as grain and void size increase linearly.

3.3.1 Porosity

The two properties of a rock or soil which are most important in controlling the behavior of subsurface water are:
(a) how much water the rock or soil can hold in empty spaces within it, and
(b) how easily and rapidly the water can flow through and out of it.

The first is defined in hydrogeology by the ***porosity*** of the rock or soil. This expresses the ratio of voids in it to its total volume. (In soil mechanics, ***void ratio***, the ratio of the volume of void space to that of the solid component, is more commonly used.) The relationship between porosity and rock and soil textures will be discussed later. In general, crystalline igneous and metamorphic rocks have low porosities unless ***secondary voids*** such as joints are produced by fracturing, for example in a fault zone or at a fold axis, or are produced by chemical erosion to give solution cavities. Porosities of terrigenous sedimentary rocks and soils may also be affected locally by fracturing, but these rocks also have ***primary voids*** called ***pores***. These are spaces left between the solid grains, and they are distributed fairly evenly throughout the body of the rock or soil when it is first formed. This intrinsic primary porosity tends to be greatest in young, poorly compacted sediments which have not been deeply buried and compressed at any stage of their geological history, and tends to be least in old, well compacted rocks buried under a thick overburden.

The rock porosity is not the only parameter which should be known, but the knowledge of other rock parameters , which have been defined in the previous chapter, is also important which could be illustrated through the relation among rock constituents (Fig. 3.1).



$W=W_a+W_w+W_s$ $V=V_a+V_w+V_s$

Fig. (3.1). The relation between rock constituents.

3.3.2 Factors Affecting Porosity of Rocks

The factors that control the porosity of sedimentary rocks and soils are as follows:

(a) The grain size variation: Since small grains can fill the voids among larger grains. A sediment with a large variation in grain size (a well graded sediment) has a lower porosity than a poorly graded sediment.

(b) The shape of the grains: Since angular laths such as occur in clay minerals often form bridges between other grains, hold them apart and increase porosity. Since porosity is dimensionless, the size of grains and voids does not affect the ratio of their volumes, and in a sediment formed of perfect uniform spheres the porosity would be independent of the size of the spheres. In practice, the different characteristics of clay minerals compared with quartz grains lead to an increase in porosity with ***reduction*** of grain size. This is because all clays adsorb water on to their outer surfaces. By this means, every clay **micelle** (tiny crystal) is surrounded by a thin film of water, which situation effectively increases the porosity of clay rocks. In crystalline limestones, the void space is mainly secondary and is controlled by the presence of fossils and bedding planes, by leaching of carbonate and redeposition by acidic ground water, and by fracturing, on both a large and small scale. Because of progressive leaching, the void space in limestone usually increases with time, and caverns may develop.

(c) The packing (or grading) of the grains: If the grains are spherical, can give a range of porosities from 26 to 47%. The looser packing is a less stable arrangement of grains (Fig. 3.2a), and a change from this to a more stable arrangement (Fig. 3.2b) will reduce porosity and may lead to the expulsion of water from a sediment.

(d) The degree of cementation: That is, to what extent pore space is replaced by cement and the extent of recrystallization at points where grains touch. Both are influenced by the age and history of burial of the rock.

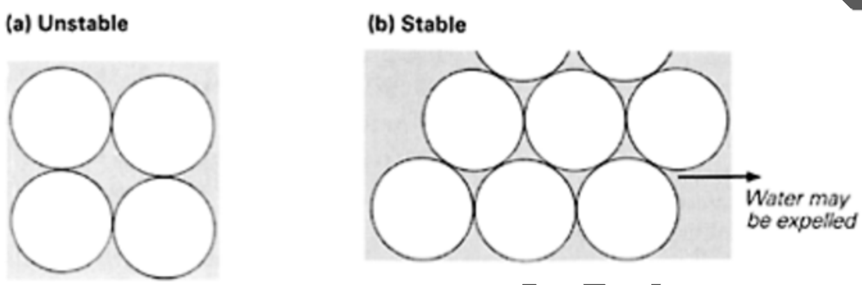


Fig. (3.2). Packing arrangements of grain in sediments: (a) unstable, (b) stable.

10.4 Groundwater, Water Table, and Vertical Classification of Groundwater

Groundwater may be divided vertically to different zones (see Fig. 3.3):

1- Zone of aeration (unsaturated zone): It is subdivided into:

a- Soil water zone (pellicular water): The first rain to infiltrate below the ground surface wets the grains of soil and adheres to them as ***pellicular water***. The forces holding the water to the grain boundaries are so strong that it is not moved further and can be detached only by evaporation or plant roots.

b- Gravitational water: Percolation to greater depths as rain continues to fall proceeds only after the soil reaches its ***field capacity*****,** at which it cannot hold any more water against the downward pull of gravity. Then, the next water to infiltrate moves over these first films of water, but does not fill the voids in the rock completely. There is still air present in the centers of the larger voids.

The ***pellicular*** and ***gravity*** water in this ***unsaturated zone*** **(**or ***zone of aeration***) is called ***vadose water***.

c- Intermediate zone: This zone may be present or absent depending on the type of ground water i.e. it is absent in areas of raising water table and its groundwater of gravitational type.

d- Capillary zone: This zone separates the aeration zone from the saturation zone. The water table lies below the top of the saturated zone in the rocks surrounding the well, separated from it by the ***capillary fringe*** (sometimes referred to as the ***capillary zone***). This is usually between a few centimeters and a few meters in thickness, but may be over 10 m thick in very fine-grained rocks in which capillary pressures are high. The smaller the voids in the rock, the higher the fringe. In the open well, the water surface is at atmospheric pressure; and at the top of the saturated zone in the rocks, water is at atmospheric minus capillary pressure. Only close to the water table are all the voids in the soil completely filled with water and, for this reason, most of the capillary fringe often lies within the zone of aeration rather than forming part of the zone of saturation in the strict sense.

2- Zone of saturation : Eventually the gravity water percolates to a ***zone of saturation***, where all the effective void space in the rock is filled with water. The journey from the surface may take up to a few weeks. The water in this saturated zone is referred to as ***groundwater***, and its upper surface is often referred to (not in strict usage of the term) as the ***water table*** (Fig. 3.3).The saturated zone persists downwards until the compaction of the rock under the pressure of overburden reduces porosity to zero. This depth varies with local geological conditions, and may be as much as 10 km in regions of thick sediments.

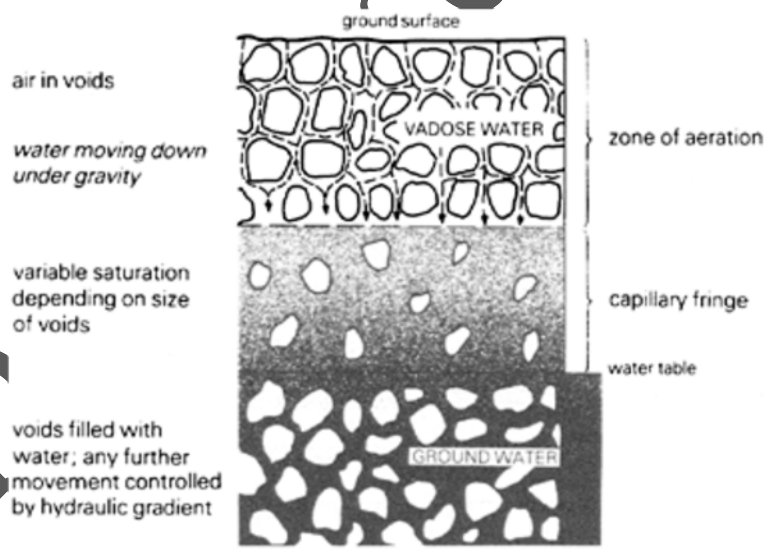


Figure (3.3). The distribution and behavior of water in rocks and soils in the subsurface.

3.5 Characteristics of Water Table

1- It is the upper surface of groundwater is often referred to (not in strict usage of the term) as the ***water table*** (Fig.3.4). The saturated zone persists downwards until the compaction of the rock under the pressure of overburden reduces porosity to zero. This depth varies with local geological conditions and seasonal changes.

2-The water table is shaped like a subdued replica of the topography above it.
3- It is not static, as groundwater in a permeable rock is continually in motion. Highs in the water become flatter, and gradients are reduced, at a rate controlled mainly by the permeability of the rock. The gradients are usually less than 1 in 100, but may be as much as 1 in 10 in hilly country. If an uncased well were drilled into the saturated zone, water would flow into it and, given time, fill it to a level which is a point on the water table, as precisely defined.
4- Definition of the water table and of its variation seasonally and over longer periods is important for groundwater supply and other practical purposes. It can be located and monitored by wells, and less accurately by ***geophysical surveys*** (such as electrical resistivity and seismic refraction techniques). In large investigations both would be used.
5- Natural discharge of groundwater takes place where the ground surface intersects the water table. If the flow from the hydrologic unit is spread diffusely over an area of marshy ground, it is usually referred to as a seepage. If it is concentrated, say by a fissure acting as a channel, it is called a ***spring***.

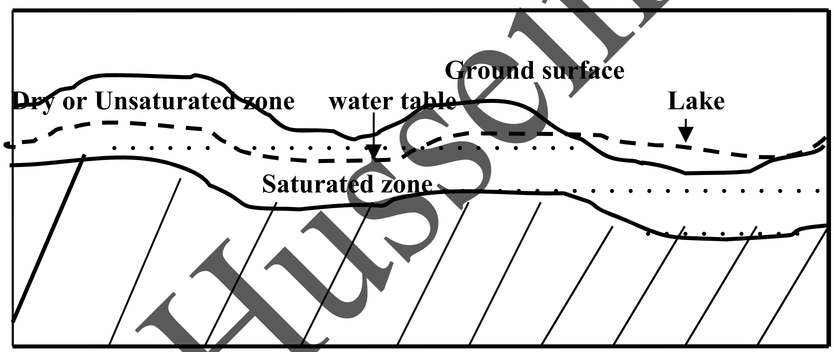


Fig. (3.4). Groundwater table and its relation to topography.

3.6 Types of Aquifers

Circulation of groundwater in the saturated zone takes place within aquifers, but the zone of movement is effectively limited by any ***aquicludes*** bordering the aquifer. Consequently, the subsurface is divided by geological structure into more or less self-contained ***hydrological units***, or ***groundwater basins***. Within each of these there is an approximate balance of supply and discharge, and flow is independent of other units. The units vary considerably in structure and in size. A hydrological unit may be a thick layer of sandstone several hundred kilometers square in area, a body of gravel in a buried channel, or an alluvial fan (Fig. 3.5) only 100 m across.

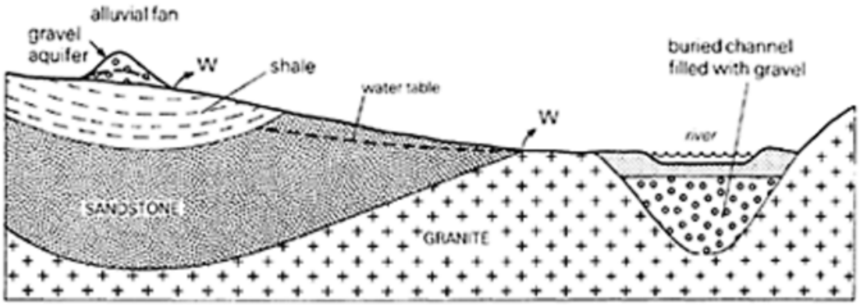


Fig. (3.5). The diagram shows three examples of a hydrological unit ,
(a) an alluvial fan with a gravel aquifer and discharge at W,
(b) a thick sandstone layer, and
(c) the gravel infill of a buried channel.

The common aquifers include: (a) sand and gravel deposits, occurring as drift in glaciated areas, or as alluvial cones (especially in semi-arid hilly regions); (b) sandstones; (c) certain permeable limestones such as chalk; and (d) fissured or deeply weathered igneous rocks, especially in tropical areas.

3.6.1 Unconfined Aquifers

An aquifer bounded only from below by an ***aquiclude*** is termed an ***unconfined aquifer*** **(**Fig. 3.6). Flow to the surface may also come from unconfined aquifers, where a well is sited near the bottom of a steep valley. The level of the water may fluctuate considerably during the course of a year, dropping during dry seasons and rising following periods of rain. Therefore, to ensure a continuous supply of water, a well should penetrate many meters below the water table. When water is pumped from a well, the rate at which water flows in from surrounding rock to replace that which is extracted is generally slower than the rate at which water is taken out. Thus, in an unconfined aquifer, it produces a conical depression in the water table around the well that is known as a ***cone of depression***. If a well is drilled into an unconfined aquifer, the water will rise in the well to the same height as the water table in the aquifer. The water must be actively pumped to the ground surface.

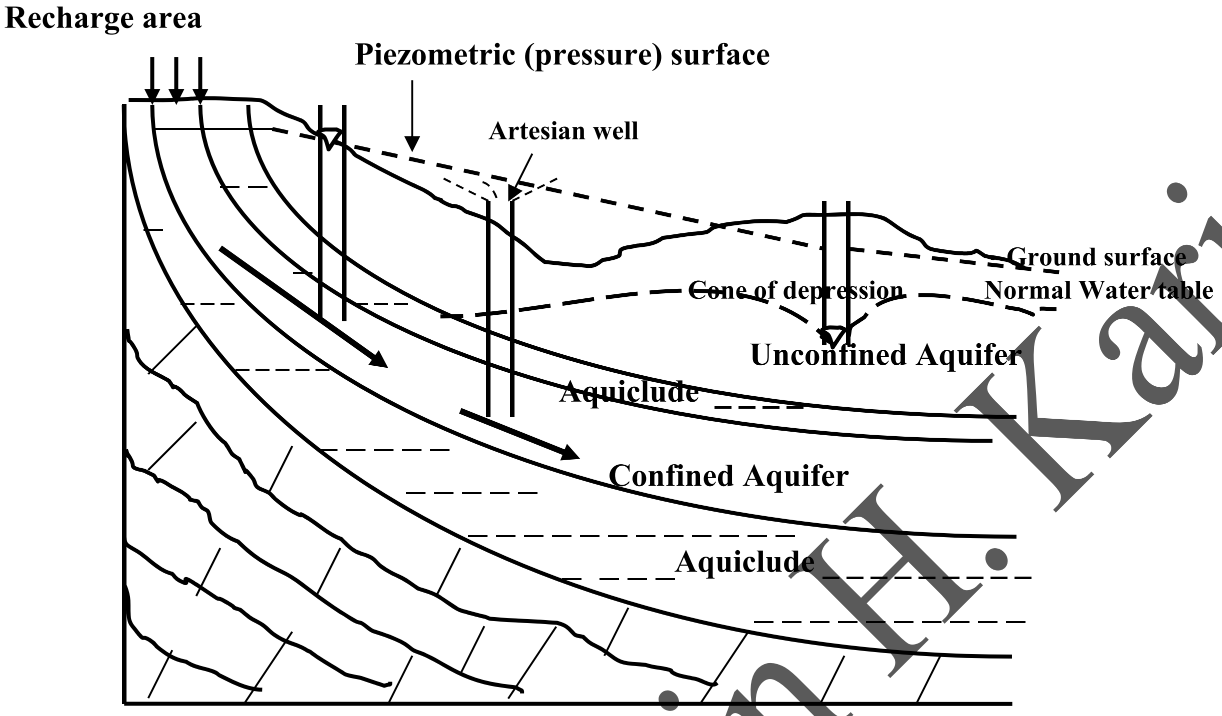


Fig. (3.6). Unconfined and confined aquifers.

The main characteristics of unconfined aquifers:

1- The unconfined aquifers are in direct contact with the atmosphere.

2- The water presented inside rocks are under the effect of atmospheric pressure alone, so these are called ***free aquifers***.

3- The level of the water in these aquifers are affected by several factors such as topography of the area, amount of falling water in the region and the exhausted water from the wells.

3.6.2 Confined (Artesian) Aquifers

When an aquifer bounded both from below and above by an aquiclude, retarding or effectively stopping any flow of water to the surface, the aquifer is said to contain ***confined water***. Water in a confined aquifer may be under considerable pressure from the adjacent rocks, or as a consequence of lateral differences in elevation within the aquifer. Figure 3.7 shows an aquifer which crops out on high ground and dips under an aquiclude. The outcrop serves as an intake area for percolation, which recharges the confined water elsewhere in the induced artificially by flow to the surface through wells. The inclination of the ***piezometric surface*** is controlled by the rate of flow through, and the permeability of the aquifer, as described by Darcy's Law, and as demonstrated by the experiment illustrating it. The piezometric surface is the surface where the hydrostatic pressure inside the aquifer is equal to the atmospheric pressure (pressure surface). If the piezometric surface is higher than the water table in a porous aquiclude, then groundwater conditions are ***artesian*** rather than normal or subnormal. If the piezometric surface lies above ground level within this ***area of artesian flow*, **then water will rise in an open tube to give a ***flowing artesian well***.

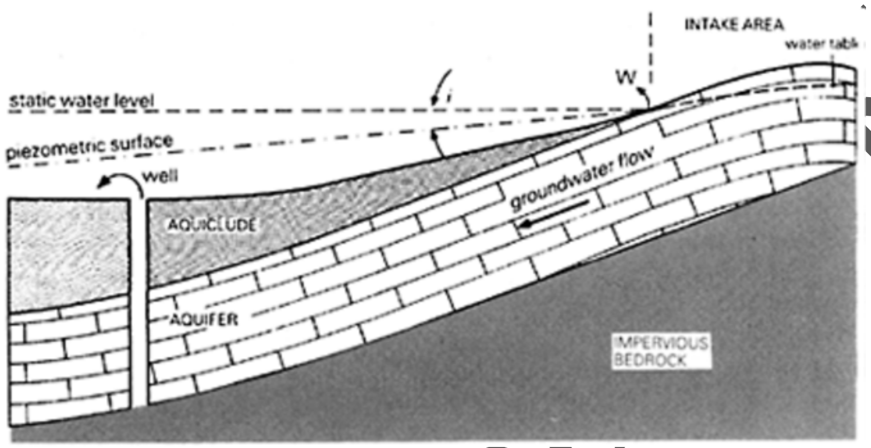


Fig. (3.7). The ground water in the aquifer is confined by the aquiclude above it, except at the intake area where it crops out, and can be recharged by infiltration. The free water table within the intake area, and the spring line *(W)* at the contact of aquifer and aquiclude, are shown.

Depending on the permeability of the aquifer and the amount of groundwater flow, the piezometric surface slopes at some angle ***i*** from the intake area. If the surface lies above ground level at the well, as shown, there is artesian pressure enough to make water flow from the well.

The basic geological conditions for artesian pressure, apart from rare cases where there is a major source of supply to the aquifer other than at the outcrop, are that groundwater be confined and that strata be inclined. These are not uncommon, and neither are artesian conditions. They may occur very locally in drift where sand or gravel lenses are covered by clays, or they may affect hundreds of square kilometers underlain by a permeable rock group.

When a well is located where the pressure surface is below the ground surface so the well is called a non artesian well. When the pressure surface is above the ground and a well is drilled into the aquifer, a flowing artesian well is created.

The main characteristics of the confined aquifers are:

1- These aquifers are subjected to pressure higher than the atmospheric pressure as they are bounded by impermeable layers, both above and below the aquifer, must be present to prevent the water from escaping.

2- Water must be confined to an aquifer that is inclined so that one end is exposed at the surface where it can receive water, that is called the ***recharge area***.
3- When such aquifers are penetrated by wells, the rise and lowering of water inside these aquifers depend upon the variations resulted from pressure, amount of stored water which are different from that in the unconfined aquifers.

10.7 Aquifer and Porosity

In order that water (or any other liquid, such as oil) can flow through rock, it is necessary that the rock not only be porous, but not permeable. That is, the pores (or at least some of them) must form, in effect, a connected network of tiny pipes through which water can move. Figure (3.8a) shows porous but impermeable rock and Figure (3.8b) shows permeable rock.

(a) **(b)**

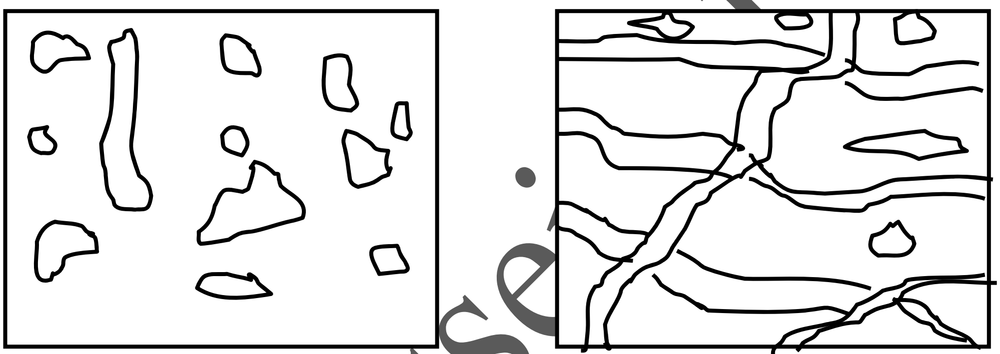


Fig. (3.8). Impermeable (a) and permeable (b) rocks.

Note that some of the pores in Figure (3.8b) do not form part of the connected pore system, but are isolated from it. Any water contained in these pores is cut off from the system of connected pores. (Evidently, this water found itself trapped in the isolated pores that formed pumped out of the rock. Rather, this water is retained by the rock).

Let V_{retain} be the total volume of the isolated pores (and hence of the retained water if the pores are filled), and V_{yield} the volume of all the connected pores (and hence the volume of the water the rock can yield if these pores are filled) in a certain block of rock. If the total pore volume V_{pores}. The volume of the block of rock , as given from its external dimension, is V.

The total pore volume V_{pores} is given by:

$$V_{pores} = V_{retain} + V_{yield} \quad (3.1)$$

dividing by V gives :

$$V_{pores} / V = V_{retain} / V + V_{yield} / V \quad (3.2)$$

Now $V_{pores} / V = n$ **,** the porosity of the rock **;** in similar manner, define the yield porosity n_y by

$n_y = V_{yield} / V$ (3.3)

and the retention porosity n_r , by

$n_r = V_{retain} / V$ (3.4)

Therefore, by :

$n = n_r + n_y$ (3.5)

The porosity that counts in groundwater flow is the yield porosity n_y, sometimes called the ***effective porosity***. The isolated pores, and any water they contain , are part of the rock matrix. As stated above, the formation, or layer, of permeable rock is called an ***aquifer***, whereas a layer of impermeable rock is an ***aquiclude***.

3.8 Darcy's Law

Slow, laminar flow of water through a porous medium within which the stream lines are smooth parallel paths, and where there is no turbulence or eddies in the wake of grains, is described by an empirical formula, Darcy's Law. This law is valid under most conditions found in nature, where the critical value of the Reynold's Number (at which turbulent flow develops) is seldom approached.

Consider an aquifer in which the water table is horizontal everywhere. In this configuration, the water will just sit in the connected pores, there is no tendency for the water to move in any horizontal direction. The reason is that gravity acts vertically down and has no horizontal component. Since the water table is horizontal, there is no component of gravity parallel to the water table. But if the water table is presumed to be inclined at an angle i with the horizontal, and it is presumed that i is not too large. (If the angle is too large , the velocity of the water may reach values so that turbulence sets in and Darcy's is not valid under such circumstances). As a result of being inclined , the water table drops a vertical distance h over a horizontal distance l or s. Darcy's law states that the Darcy's velocity in its simplest expression

$v_D = K(h/ l)$; or (3.6)

$v_D = Ki$ (3.7)

where v_D , the Darcy's velocity , is the volume of water ***(Q)*** discharged through a unit area ***(A)*** of a porous medium in which the pores are saturated, and inflow rate to the medium equals outflow rate, that is:

$v_D = Q/A$

$Q = A.\ v_D$ (3.8)

The actual velocity v of an aquifer calculated by dividing Darcy's velocity v_D by the effective velocity n_y:

$\boldsymbol{v = v_D / n_y}$; or
$\boldsymbol{v_D = v\, n_y}$ (3.9)

The quantity ***K*** is called the ***hydraulic conductivity*** and sometimes called the ***coefficient of permeability***, which is often very small so its unit is in m/d instead of m/s. Since the units of ***h*** and ***l*** in Eq.(3.6) cancel, ***K*** is seen to have the same units as $\boldsymbol{v_D}$, i.e., m/s in SI units. The numerical value of ***K*** depends on the properties of the rock and on properties of the liquid. Specifically, for the rock it is the degree to which the pores are connected, and their size, that enters into ***K***. For the liquid, the viscosity (internal friction) and density affect the value of ***K***. This means that for any one given kind of rock, there are different values of ***K*** for water and for oil, for example. ***i*** is the ***hydraulic gradient***, that is, the rate, ***dh/dl*** or ***dh/ds,*** at which the hydraulic or pressure head of water, ***h,*** measured from a horizontal datum, changes laterally with horizontal distance, ***l*** (or ***s***).

Now, the movement of water through rock is usually a very slow process. The pores are very small , their geometry complex, and friction ever present. This means that the real flow velocity ***v*** is usually very small. For this reason, hydraulic engineers often use the day (abbreviation d) rather than the seconds (s) as a unit of time. That is, values of ***v*** , $\boldsymbol{v_D}$ and ***K*** are generally expressed in m/d rather than the base units of m/s. For water, values of ***K*** for various rocks range between near zero and several hundreds of meters per day.

Equation (3.6) can be combined with Eq.(3.8) to yield this equation for the discharge:

$\boldsymbol{Q = A.\ K\ (h/\ l)}$; (3.10)
Some times this equation, rather than Eq.(10.6), is referred to as ***Darcy's law***.

Darcy's law is sometimes expressed in terms of the angle ***i*** of inclination of the water table. Since ***i*** is small (for Darcy's law to be valid), $\boldsymbol{i \approx \tan i}$ when ***i*** is expressed in radians. From Figs. (3.9 and 3.10) notice that $\boldsymbol{\tan i = h/\ l}$. Hence, for small ***i***, $\boldsymbol{i = h/l}$. Equation (3.10) can now be written as:
$\boldsymbol{Q = A.\ K.\ i}$ (3.11)
Sometimes, this is called ***Darcy's Law***.

The physical condition of the ***hydraulic gradient , i (=dh/dl)*** , may be illustrated experimentally. The pressure head at a given point within a body of rock or soil which is saturated with water is the level to which water would rise if an open-ended tube were inserted to tap the hydraulic pressure at that point. A tube that serves this physical purpose in a laboratory experiment is called a piezometer. In geological conditions where the subsurface water in a permeable rock is sealed and confined below an impermeable layer , the level to which water rises in a cased well penetrating the permeable rock is usually referred to as the piezometric level rather than the pressure head. Piezometric levels are points on a piezometric surface, the maximum tilt ***(i)*** of which from the horizontal defines the hydraulic gradient ***dh/dl*** in the water saturating the confined permeable rock. The physical meaning of these terms and Darcy's Law may be demonstrated by the apparatus shown in Figure 3.9. This consists of a reservoir (A) joined to a horizontal conduit (B) through which water can flow from A to a tap (C). Six open-ended tubes (manometers) rise vertically from B. The pipe B is filled with soil of uniform permeability k_1 . The tap is closed and the reservoir A is filled until water reaches the level h_0, and remains there after water has percolated through the soil to C and risen in each of the tubes to the static water level, h_0. When the tap is opened and the reservoir is replenished continuously to maintain the level at h_0 within it, a new equilibrium is attained eventually for steady flow through the soil, and for steady discharge v at C. Under these steady-state conditions the levels in the tubes fall to h_1 , h_2 , and so on. Each tube measures the pressure head at the point where it joins the conduit B. There is a progressive regular drop in head ***(dh)*** with horizontal distance ***(dl)*** from A to C, giving a steady inclination $i=dh/dl$ of the piezometric "surface". (The experiment is limited to a one-dimensional traverse of a permeable medium). If the soil k_1 between A and the midpoint (M) of the conduit were replaced by a less permeable k_2 soil, then, under steady-state conditions of flow with the reservoir level at h_0, there would be a reduction in the bulk hydraulic conductivity of the soils in B, in the rate of flow through them, and in the amount of water discharged at C. The section of the piezometric surface between A and C would have a deflection point at M, with the inclination i_2 of one segment corresponding to the hydraulic gradient through k_2 , and the lesser gradient i_1 to k_1. The relevance of this experiment to natural phenomena, particularly the flow of groundwater and the pressure of pore water, is discussed as they are described.

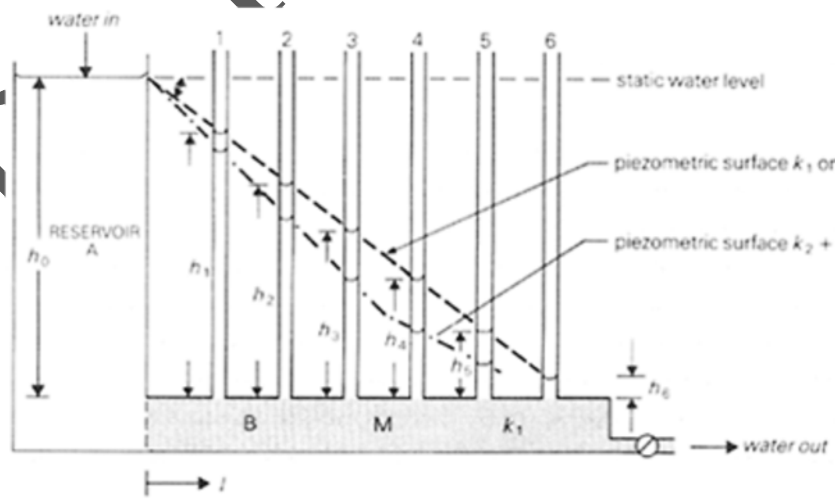


Fig. (3.9). An experiment to illustrate the relationships between permeability, change of head with distance and inclination *(i)* of the piezometric surface. A description and explanation are given in the text.

3.8.1 Unconfined Aquifer Case

Figure (3.10) shows an idealized setting of a typical unconfined aquifer. The uphill source of the groundwater, perpendicular over a certain land area, is not shown. The ground surface slopes gently, i.e. at a small angle ***i***, from left to right. Usually, the water table is roughly parallel to the ground surface, falling a vertical distance ***h*** over a horizontal distance ***l***. Here, the additional assumption is made that the layer of impermeable rock beneath the aquifer also slopes down at an angle ***i***. This means the vertical thickness ***h*** of the water-bearing part of the aquifer has everywhere the same value. Two wells have been driven into the aquifer, piercing the water table at points A and B (Fig. 3.10). The time ***t*** needed for the groundwater to move from A to B . The time t needed for the groundwater to move from A to B can be measured by dumping marked water into the well at A and noting when it shows up at well B. Strictly speaking the travel time from A to B is given by:

$t = AB / v$

Now

$AB = (l^2 + h^2)^{1/2}$

Since ***i*** is small, $h^2 << l^2$, so that AB ≈ ***l. This approximation is appropriate not only because the hydraulic*** gradient is small, but because the values of the hydraulic conductivity and effective porosity are not likely to be known with great precision throughout the aquifer. Therefore, it is sufficiently accurate to use:

$t = l / v$ (3.12)

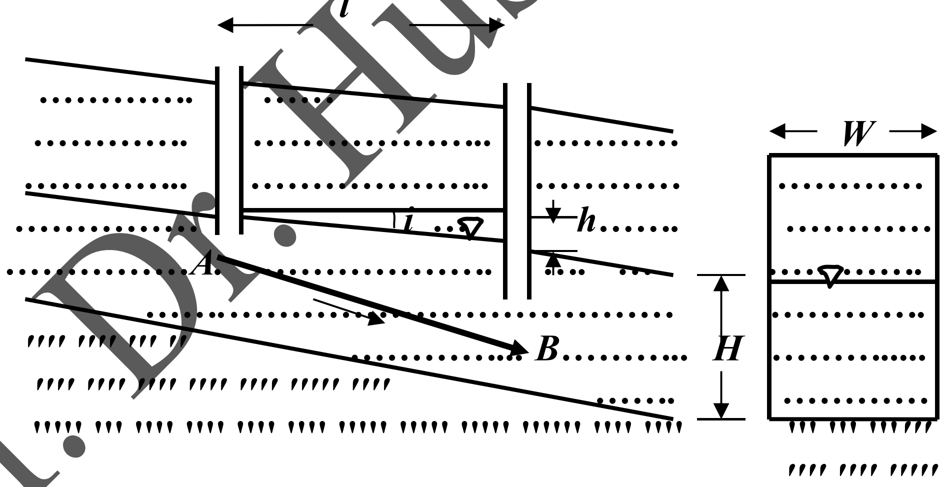


Fig. (3.10). Setting of an unconfined aquifer .

EXAMPLE 3.1: An aquifer with cross-section (horizontal width) 265 m and a vertical thickness under the water table of 42 m. The water table is 36 m below the ground surface. The discharge at this section is 3340 m^3/d and the yield porosity 27.1%.

It is required to find:

(a) Darcy’s velocity. (b) Actual velocity of aquifer water.

(a) Calculate aquifer cross-sectional area
$\boldsymbol{A = W. h} = (265\text{m})\ (42\text{m}) = 1.113*10^4\ \text{m}^2$
$\boldsymbol{Q = A\ v_D}$
$0.03866\ \text{m}^3/\text{s} = (1.113*10^4\ \text{m}^2)\ \boldsymbol{v_D}$
$\boldsymbol{v_D} = 3.47 * 10^{-6}\ \text{m/s}$
(b) Calculate the actual velocity: Darcy's velocity $\boldsymbol{v_D}$ is converted to actual velocity by means of yield porosity (substituting with fraction)
$\boldsymbol{v_D = n_y . v}$
$3.47 * 10^{-6}\ \text{m/s} = \text{o}.271\ \boldsymbol{v}$
$\boldsymbol{v} = 1.28* 10^{-5}\ \text{m/s}$
$\boldsymbol{v} = 0.0128\ \text{mm/s}$

EXAMPLE 3.2: The longitudinal and transverse cross-sections of an unconfined aquifer are shown in Fig, (3.11). Water takes 1.91 y to move from well A to well B. The hydraulic conductivity of the aquifer rock is 135 m/d. (a) Find the yield porosity of the aquifer rock. (b) It is found that $8.42 * 10^5\ \text{m}^3$ of water passes through any cross-section of the aquifer in 2.00 weeks. Find the width ***W*** of the aquifer.

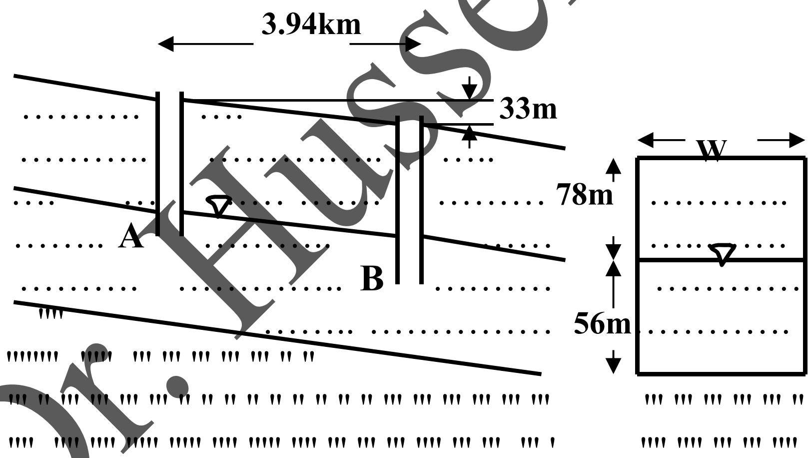


Fig. (3.11) . Example 3.2.

(a) Reading distances from Fig. (10.11), the data are: $\boldsymbol{l}$ = 3940 m, $\boldsymbol{h}$=33 m, (not 78 m: the water table is parallel to the ground surface, and therefore drops by the same distance as does the ground surface); H=56 m (thickness of aquifer below the water table). Also $\boldsymbol{t}$ = (1.91 y) (365025 d / y), $\boldsymbol{t}$ = 697.6 d ; $\boldsymbol{K}$ = 135 m / d . By Eq. (10.6), Darcy's velocity $\boldsymbol{v_D}$ is:
$\boldsymbol{v_D = K(h/ l)}$
$\boldsymbol{v_D} = (135\ \text{m/d})\ (33\text{m} / 3940\ \text{m})$
$\boldsymbol{v_D} = 1.131\ \text{m/d}$
The actual velocity of the water follows from Eq.**(**3.12**) :**

$v = l / t$
v = 3940 m / 697.6 d
v = 5.648 m / d
By using Eq. (10.9) :
$v_D = n_y . v$
1.131 m / d = n_y (5.648 m / d)
n_y = 0.200 (20.0 %)
(b) Evidently , the discharge is :
$Q = v / t$
Q = (8.42 * 10^5 m^3 / 14.0 d)
Q = 6.014* 10^4 m^3 / d
Note that the symbol t for time in this part is not the time for the water to move from A to B. It would be tedious to use two different, or subscripted, symbols for various time intervals. As usual in such situations, the context of the discussion reveals the appropriate identification for t. Now apply Eq. (3.8) with $A = w.h$, so that:
$Q = w.h\ v_D$
6.014* 10^4 m^3 / d = w (56 m) (1.131 m / d)
w = 950 m

EXAMPLE 3.3: The aquifer shown in Fig. (3.12) has a width of 7.10 km. Industrial waste BXC is illegally dumped into well A, near the factory. The material moves with the ground water , and appears at well B after 6.40 y. The effective porosity of the aquifer rock is 0.350. Calculate: (a) the hydraulic conductivity of the aquifer rock and; (b) the aquifer discharge.

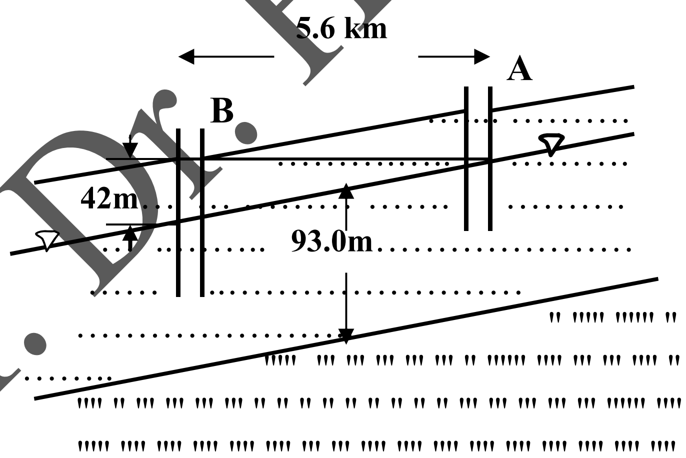


Fig. (3.12). Example 3.3.

$v = l / t$
v = (5.6 km) / 6.4 y)
v = (5600 m) / (2337.6 d)
v = 2.3956 m / d

$v_D = n_y \cdot v$
v_D = (0.35) (2.3956 m / d)
v_D = 0.83846 m / d
$v_D = K (h / l)$
0.83846 m / d = K (42m / 5600 m)
K = 112 m / d
$Q = (w.h)\, v_D$
Q = (93 m)(7100 m) (0.83846 m / d)
$Q = 5.54 * 10^5\ m^3 / d$

3.9 Effect of Type and Nature of Rocks on Groundwater

Groundwater are stored in the openings occurred in rocks. Briefly they are presented in three main types:
1- Groundwater are found in the pores between and around grain particles of rocks and in the surface crust of the earth.
2- Groundwater are found in fissures, joints and fractures in the rock masses.
3- Groundwater are found in small-large cavities and caves and places of volcanic eruption.

As mentioned above, the main characteristics of rocks that affect the groundwater flow are porosity and permeability. Moreover, several factors affect the two properties. We will discuss the effect of the presented regular openings (i.e. pores) in the sediments and sedimentary rocks (Fig. 3.13).

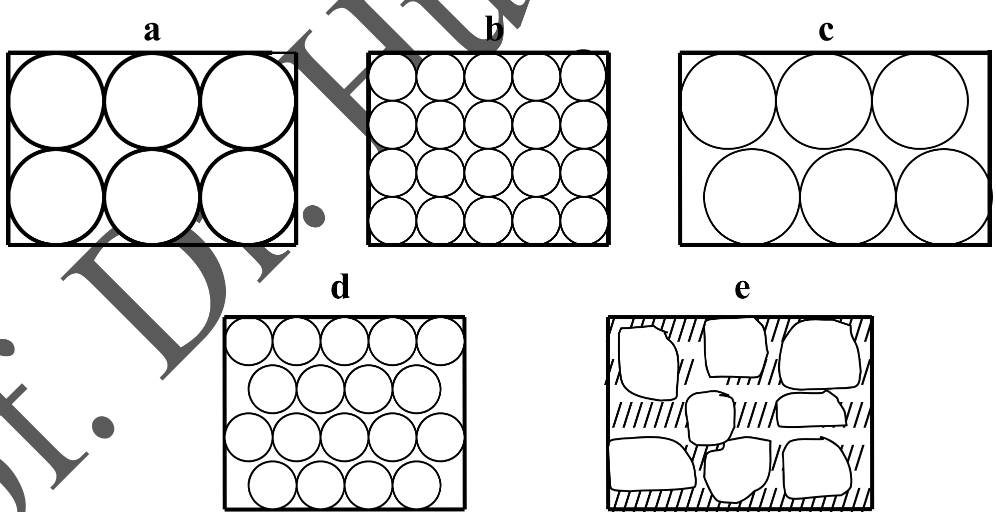


Fig. (3.13). The relation between porosity, compaction and sorting.

From the above Figure (3.13a and b), it is noticed that both sections consist of large and small balls compacted in the same way and hence both have similar porosity. This is due to constant compaction and consequently the porosity remains constant too. But the permeability is higher in the larger balls case than in the smaller one , and this is due to larger openings in the first case allowing liquids to pass easily and quickly .Both sections are well sorted. Considering Figure (3.13c and d), it seems that the distance between balls becomes closer than that in sections a and b and with the same sorting but with less porosity. If any cemented solid materials are presented among particles, and also if any smaller particles are presented between the larger ones (Fig.3.13e), the porosity decreases.

In case of clays that have high porosity but with very low permeability, it seems that the above discussion is not agreed, this is due to very small spacing between particles not allowing liquids to pass through them. Thus, the saturated clays are not considered as aquifers but they are aquicludes. Moreover, the unsaturated rock layers with high porosity and permeability are also not considered as aquifers. Consequently, good aquifers are those consisting of clean, large gravels or coarse, well sorted sands or mixture of alluvial deposits consist mainly silts. Good rock aquifers are mainly sandstones, fractured limestones contained dissolution channels and fragmented volcanic rocks.

3.10 Types of Groundwater Movements

We know that the flow of fluids is in two ways:

1- Laminar (Lamellar) Flow: The flow moves as if it were composed of lamellae, or sheets, each moving smoothly past the other with no interchange of fluid between them. This flow is occurred when the water movement is very low. as in groundwater and this is called seepage. The rate of seepage is directly proportional with hydraulic gradient and inversely proportional with permeability.

2- Turbulent Flow: When the fluids moves with an extremely complex and irregular twisting, or eddying motion, and upwelling currents.

Whether flow is laminar or turbulent, the groundwater movement depends primarily upon the permeability of rock layers and hydraulic gradient which is controlled by ***Darcy's Law***.

3.11 Effect of Geological Structures on Groundwater

The geological structures have a great effect on groundwater ant its water table level, of these structures are:

1- The presence of permeable layer overlying horizontal impermeable layer: In this case, the groundwater is blocked over the impermeable layer. The water table follows the topography of the area and its levels varied with seasonal changes (Fig.3.14).

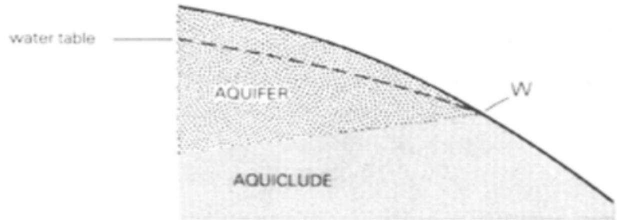


Fig. (3.14). The presence of permeable layer overlying impermeable layer.

2- The presence of small impermeable layers (lenses or pockets) within permeable layer: In this case, for each impermeable layer has its own isolated water level with different depths such as perched water table shown in Figure 3.15.

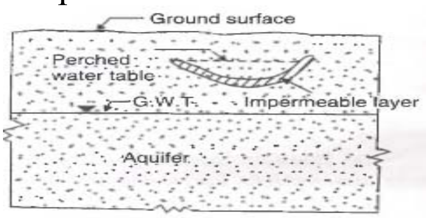


Fig. (3.15). The presence of isolated water levels (Perched water table).

3- The presence of impermeable inclined rock layer (such as dike) between two permeable layers: In this case, the impermeable inclined layer separate water on both sides and each side has its own water table at different depth (Fig. 3.16).

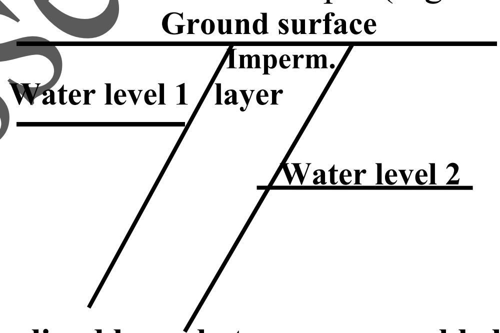


Fig.(3.16).The presence of impermeable inclined layer between permeable layers.

4- The presence of basinal structure: In this case, the water tables formed on both sides of the impermeable layers (Fig.3.17 and 3.18). The isolated water levels are not linked with the general water table in the region, These water levels are affected by climatic changes.

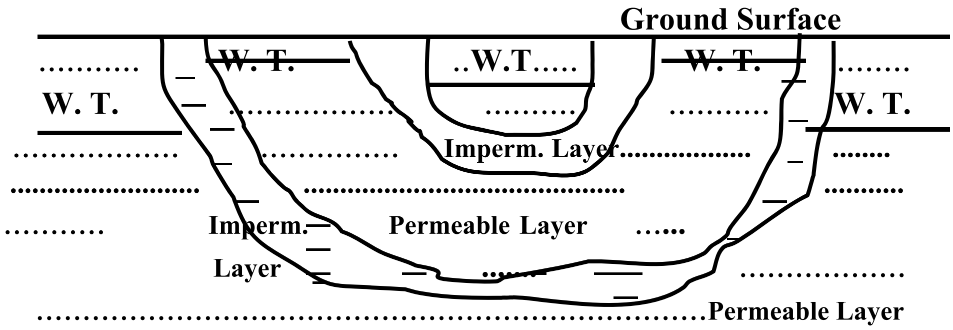


Fig. (3.17). The presence of basinal structure.

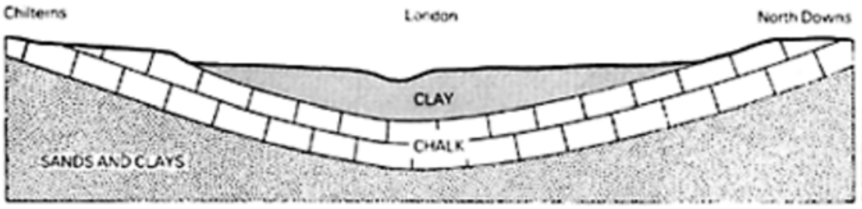


Fig. (3.18). A simplified geological section across the London Basin, which shows the chalk aquifer confined by the impermeable London Clay.

3.12 Springs

Natural discharge of groundwater takes place where the ground surface intersects the water table. If the flow from the hydrologic unit is spread diffusely over an area of marshy ground, it is usually referred to as a ***seepage***. If it is concentrated, say by a fissure acting as a channel, it is called a ***spring***. Bryan subdivided springs into:

a- Non-gravitational Springs: They include volcanic springs (***Geysers***)and those resulted from joints, fissures and fractures located at a great depths in the earth crust. In limestone areas, groundwater usually follows channels along bedding planes or fissures, which are progressively widened by solution, and discharge from limestone is often from ***solution channel springs***. The major discharge of fresh groundwater takes place at the sea coast. Most of this type are with high temperature and sometimes highly mineralized containing sulfur (***sulfur-rich springs***), such as ***Ain- Kibrit*** in Mosul City and ***Hamam Al-Alil*** in Ninawa Governorate in Iraq.

b- Gravitational Springs: In this type, the water flows due to the effect of hydrostatic pressure and subdivided into:

1- **A valley spring:**(Fig. 3.18a) occurs where the water table intersects the bottom of a valley that is cut into pervious rocks. If the water table rises and falls seasonally, flow is intermittent, and the spring is called a ***bourne***. These are particularly common in the chalk lands of southern England.

2- **Contact spring:** If the intersection of surface and water table is controlled by geological structure, and discharge takes place where impervious rock bounds the hydrologic unit (Fig. 3.18b), the overflow is called a ***contact spring*, **or sometimes a ***stratum spring*** if the rock is layered.

3- **Spring line**: The impervious barrier may be a layer of clay or shale, or an igneous intrusion. By converse argument, a ***spring line*** known to be related to a particular geological contact may be used to trace its outcrop laterally when mapping.

4- **Fault spring**: Fault springs occur where pervious rocks are faulted against impervious rocks (Fig. 3.18c), and may be thought of as a type of contact spring. If the rock in the fault zone is a pervious breccia, a natural artesian flow may bring confined water to the surface from a concealed aquifer. In desert areas this flow may produce an oasis.

5-Artisian Springs: As mentioned above, they are resulted from groundwater rises in a well above the level where it was initially encountered.

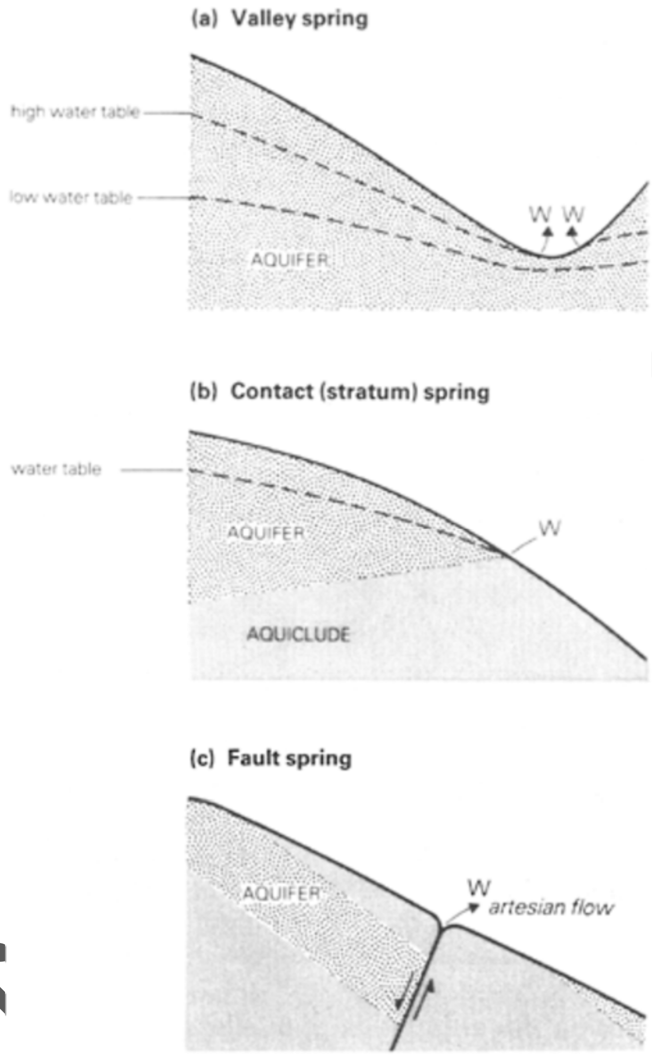


Fig. (3.18). Springs and seepages.

(a) The valley is cut into uniformly permeable rock (for example, chalk) and valley springs occur where the water table intersects the ground surface. A fall in the water table after drought lowers and flattens it, and may result in the springs drying up.

(b) A stratum spring sited at the contact of an aquifer and an aquiclude.

(c) Flow from a confined aquifer is taking place through the breccia in the fault zone, and produces a fault spring.

3.13 Quality of Groundwater

Groundwater may be subdivided according to the dissolved salts:

1- Fresh Water: Waters in which the total dissolved solids are less than 1 gm/l. These waters are used for drinking ,irrigation and for industrial purposes, such as water founded in Bukhtiari rocks .

2- Brackish Water: The amount of dissolved salts is between 1-10 gm/l. It is used mainly for irrigation, such as water in the Eocene and Pliocene limestone in western desert of Iraq.

3- Saline Water: The amount of dissolved salts is between 10-50 gm / l, such as water founded in Al-Fatha formation of the Miocene age.

4- Brine Water: The amount of dissolved salts is greater than 50 gm / l, such as water founded in the Quaternary rocks in Iraq.

3.14 Hydrogeological Subdivisions of Iraq

From the hydrogeological point of view, Iraq may be subdivided into three main regions:

1- Northern and North Eastern Region: The water within this region is of high quality and high yield with salinity between 100- 1000 ppm, such as Erbil, Makhmor, Badra and Jassan, Alton Kopri, and Tuz basins.

2- **Middle Region**: It includes all areas between Tigris and Euphrates rivers. Water in the upper part of this region (such as, Sinjar area) is of medium quality with salinity 500-4000 ppm, with intermediate quantity which is founded mainly in Injana (U. Fars) formation. While, water in Fatha (L. Fars) formation is with bad quality and less quantity. In northern Baghdad, it is found also with bad quality and less quantity.

3- South Western Region: It includes western desert in Iraq and subdivided into the following basins:

- **a- Al-Hammar and N. Qa'ara Basins:** Their water is with a salinity of 500-3000 ppm with medium yield and mainly used for drinking water , irrigation and industrial purposes.
- **b- Bahr Al- Najaf Basin:** It is with high salinity between 2000-3000 ppm with limited yield and not used for most purposes.
- **c- Al-Zubair Basin:** It consists of two different water bearing layers , one of them (the upper) is of fresh water which is lying over a saline water layer (the lower). The main discharge of the upper layer is from raining water, so they are with limited quantity and very brine water with a salinity exceeding 8000 ppm.

3.15 Investigation of Groundwater

Groundwater may be investigated by the followings:

a- Geophysical methods: This is mainly done:

1- Seismic refraction and;

2- Electrical resistivity methods.

b- Geological methods: It include the study of porosity, permeability of rock layers and the geological structures that affect aquifers and superperposition of the layers, in addition to detect layers with high probability to be aquifers.

c- Subsurface methods: They include:

1- Tested wells for monitoring.

2- Logging: It includes electrical, self potential, thermal acoustic and radioactive logging.

3.16 Groundwater and Engineering

The presence of water in rocks and its movement through them are of great importance to many human activities such as farming. However, this brief account is limited to some of the more important geological aspects of groundwater which are linked to engineering. These are: (a) water supply from subsurface sources, (b) drainage of marshes, and (c) disposal of toxic waste underground. The other aspects of equal or greater importance to most engineers, which are normally covered in a separate course of soil mechanics, are omitted or have been referred to briefly. They include pore fluid pressure in soils, the effect that the presence of water has on the strengths and other mechanical properties of soils, and the relationship between settlement of buildings and the loss of pore water from clay.

3.16.1 Groundwater Inventory

In arid regions, groundwater may be the only dependable source of water supply for much of the year, unless water is piped over long distances. It has the bonus quality of being filtered naturally. There are advantages also, which may outweigh any extra cost per liter, in having some supply from wells in areas like England that depend mainly on surface reservoirs. Problems such as creating a watertight reservoir in chalk lands, or acquiring its site, are avoided. The water is often more palatable than alternative surface water. Underground reservoirs are seldom drained entirely during droughts, and deep wells can draw on reserves temporarily, on the understanding that replenishment will be allowed to take place during the next winter. An important consideration in many schemes of supply is that groundwater resources can be developed more rapidly than a surface reservoir can be built. The lead time between the conception of a reservoir and its completion is conditioned not only by construction times and technical delays, but also by the statutory procedures of acquiring approval, permissions and land, and allowing appeals by those affected. Where a relatively quick solution must be found to a supply of water or to insure against drought, a groundwater scheme should be considered.

In assessing the feasibility of any scheme of groundwater supply and in planning it, geological and hydrological information is essential. A first step is to evaluate the permeabilities of the rocks and soils present in the area, and to determine their structure. This identifies aquifers and aquicludes and outlines the hydrologic units present. The common aquifers include:
(a) sand and gravel deposits, occurring as drift in glaciated areas, or as alluvial cones (especially in semi-arid hilly regions);
(b) sandstones;
(c) certain permeable limestones such as chalk; and
(d) fissured or deeply weathered igneous rocks, especially in tropical areas.

A small local supply may be obtained from a single well in an alluvial cone, but a large scheme usually draws on the groundwater in an extensive, thick, permeable layer.

For the above reason, the prime geological factor in choosing the precise site for a boring is the probability that the permeability of the aquifer is higher at that locality, or that flow is assisted by fracturing. In soils, a lens of gravel may expedite flow from a larger body of sand; in solid rock the yield from a well sited on a fold axis, where jointing is well developed, may be several times that from massive chalk less than 100 m away. Since rocks weakened by fracturing are often eroded preferentially into hollows, the water table is likely to be at shallower depth also. ***Electrical methods*** of geophysical surveying are used to locate such water-filled fractures. In ***water table wells*,** water is pumped from an unconfined aquifer, whereas in ***confined water wells*,** the whole aquifer penetrated by the well is saturated and there is usually artesian pressure.

REVIEW QUESTIONS AND PROBLEMS

3.1 Define groundwater and relate it to the water table.

3.2 How can we recognize a confined aquifer from unconfined one?

3.3 Under what circumstances, Darcy's law is applied?

3.4 What are the rock types that contribute to salt concentrations in groundwater?

3.5 Show the relation between aquifers and porosity?

3.6 Classify rocks according to their porosity and permeability?

3.7 What are the main factors affecting rocks porosity?

3.8 How do porosity and permeability differ?

3.9 Under what circumstances can a material have a high porosity but not be a good aquifer?

3.10 Distinguish between an aquiclude and an aquifer?

3.11 What is meant by the term artesian? Under what circumstances do artesian wells form?

3.12 Describe two ways in which sinkholes are created.

3.13 Areas whose landscapes are largely a reflection of the erosional work of groundwater are said to exhibit what kind of topography?

3.14 When an aquiclude is situated above the main water table, a localized saturated zone may be created. What term is applied to such a situation?

3.15 Show the effect of geological structures on groundwater? Give examples.

3.16 Distinguish between hydraulic conductivity and coefficient of permeability?

3.17 Give the vertical distribution of groundwater?

3.18 Show the relation between groundwater level and topography?

3.19 What are the main types of porosity?

3.20 What are the main characteristics of confined and unconfined aquifers?

3.21 Comment on the water quality for drinking purposes.

3.22 Give a classification for Iraq from the hydrogeological point of view.

3.23 Give an estimation of groundwater quality.

3.24 What is the source of heat for most hot springs and geysers ? How is this reflected in the distribution of these features?

3.25 What are the main types of springs?

3.26 What are the main methods used for groundwater investigation?

3.27 how the relation between groundwater and civil engineering.

3.28 A rock has a porosity of 43.6%. About 35% of the pores are isolated. Calculate the yield porosity of the rock. **(Ans. 28.3%)**

3.29 The aquifer of Figure (3.19) has a yield porosity of 34% , a retention porosity of 19.2% and a hydraulic conductivity of 160 m/d.
a- How long time does it take to move from well A to well B?
b- What volume of water passes any cross section in 10.0 days?
c- Assuming that all pores below the water table are filled, find the total volume of water trapped in the rock between the wells.
(Ans. a- 812 d ; b- $3.16*10^5 m^3$; c- $1.45*10^7 m^3$)

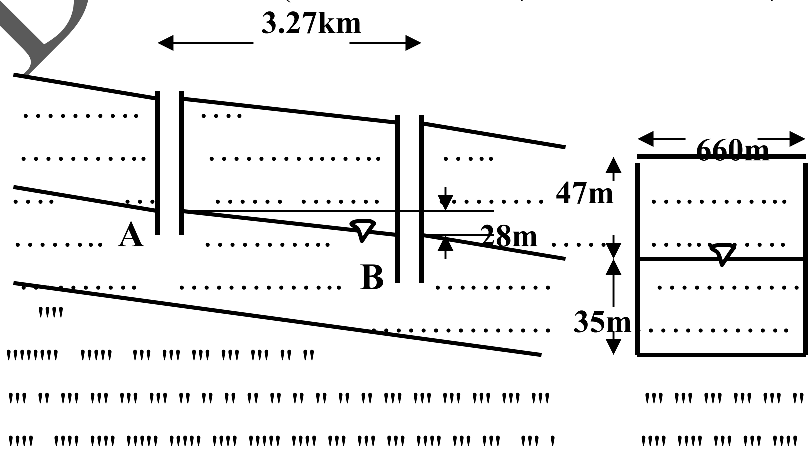


Fig. (3.19). Problem 3.29.

3.30 Calculate the Darcy speed through an unconfined aquifer 870 m wide and 106 m thick when the discharge equals 2.19 m^3/s and the water table is 18.0 m beneath the ground.

3.31 An aquifer in a certain region consists of coarse sand with a yield porosity of 39.0% and a hydraulic conductivity of 45.0 m/d. The width of the aquifer is 2.90 km and its vertical thickness below the water table is 73.0 m. The aquifer discharge equals 8380 m^3/d. Calculate the vertical drop of the water table over a horizontal distance of 6.60 km.

3.32 Over a horizontal distance of 2.87 km, the water table falls by 41.5 m. Groundwater takes 3.18 years to move the 2.87 km. The effective porosity of the aquifer is 0.640, and the aquifer' s cross - sectional area is 4500 m^2. Find the hydraulic conductivity of the aquifer rocks .

(Ans. 109 m/d)

3.33 The aquifer shown in Figure (3.20) is 1.3 km wide. The effective porosity is 0.290 and the hydraulic conductivity equals 83.0 m/d. Water containing a dangerous radioactive isotope of plutonium is dumped (accidentally, of course) into well A. How long does it take for the plutonium to show up in well B ? Assume that the plutonium is carried along at the same speed as the groundwater.

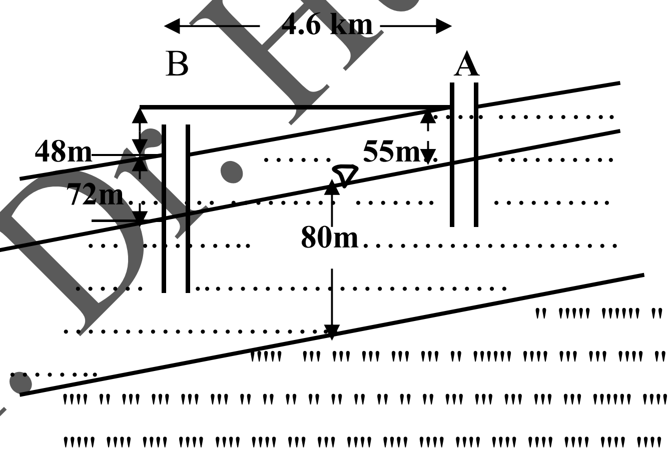


Fig. (3.20). Problem 3.33.

4. Site Investigations and Geophysical Techniques

4.1 Site Investigations

4.1.1 Introduction

Site investigations or ***subsurface explorations*** are done for obtaining the information about subsurface conditions at the site of proposed construction. A complete understanding of the characteristic features of a site are essential prior to embarking on a civil engineering construction project. The main objectives of site investigations are as follows:

1- To assess the suitability of the site for a given engineering structure.
2- For obtaining information about the surface and subsurface features which is essential to select suitable-economic design and for planning construction techniques at the proposed site.
3- Understanding the main difficulties resulted from geological conditions of the area which may be faced during or after construction and to plan and design appropriate foundations.
4- To investigate the safety of the existing structures and to suggest the remedial measures.
5- And finally, to draw up bills of quantity for excavation normally requires that most of the following information be obtained:

a- What rocks and soils are present, including the sequence of strata, the nature and thicknesses of superficial deposits and the presence of igneous intrusions;
b- How these rocks are distributed over , and under, the site (that is, their structure);
c- The frequency and orientation of joints in the different bodies of rock and the location of any faults;
d- The presence and extent of any weathering of the rocks, and particularly of any soluble rocks such as limestone;
e- The groundwater conditions, including the position of the water table , and whether the groundwater contains noxious material in solution, such as sulphates, which may affect cement with which it comes in contact;
f- The presence of economic deposits which may have been extracted by mining or quarrying, to leave concealed voids or disturbed ground; and
g- The suitability of local rocks and soils, especially those to be excavated, as construction materials. Special information such as the ***seismicity*** of the region or the pattern of sediment movement in an estuary may also be required.

4.1.2 Stages of Site Investigations

Site investigations in one form or the other is generally required for every big engineering project . Subsurface exploration are generally carried out in three stages:

1- Site Reconnaissance: It is the first step in subsurface exploration programme. It includes the followings:

a- Studying the maps and other relevant records , published information and other existing data.
b- A reconnaissance visit to the site.
c- Preparing an approximate geological maps.
d- Collecting samples from drilled testing wells.
e- Applying geophysical surveys as a reconnaissance of the subsurface geology.
f- It helps in deciding future programme of site investigations, scope of work, methods of exploration to be adopted, types of samples to be taken and the laboratory testing and in-situ testing.

In minor projects, or where the site is small and has previously been built on, only this stage may be necessary, which is cheap and always worth doing.

2- Preliminary Investigation: The aim of this stage is to establish a complete understanding of the geological structure for the site. Besides, determine the depth, thickness, extent and composition of each soil stratum at the site. The depth of the bed rock and the groundwater table is also determined. In this stage, the following purposes may be done:

a- Preparing a detailed ***geotechnical maps***: These are geological maps with engineering expressions prepared with respect to the purpose, i.e. for dam or tunnel sites.
b- The preliminary explorations are generally in the form of a few borings or test pits to obtain information about the strength and compressibility of soils.
c- Studying the results of the experimental and in-situ (field) testing of the soils.
d- Geophysical methods are also used in preliminary explorations for locating the boundaries of different strata.

3- Detailed Investigations: The purpose of the detailed explorations is to provide confirmation of the previous results by detailed boring, drilling and excavation at critical points on the site, thus any changes in the design may be done in this stage. The main objectives of this stage:

a- To ascertain the geological conditions at the project site during drilling in order to select the type and depth of foundation for a given structure.

b- Determining the engineering properties of the soils in different strata.
c- It includes an extensive boring programme, sampling and testing of the samples in a laboratory.
d- To establish the groundwater level and to determine the properties of water.

For complex projects involving heavy structures, such as bridges, dams, multi-story buildings, it is essential to have detailed explorations. However, for small projects , especially at sites where the strata are uniform, detailed investigations may not be required. The design of such projects is generally based on the data collected during reconnaissance and preliminary explorations.

4.2 Applied Geophysical Techniques

4.2.1 Introduction

Geophysics is the study of the physical processes active within the earth, and of the physical properties of the rocks forming it. From it stemmed ***applied geophysics*****,** a set of exploration methods used to infer the distribution of rocks underground from physical measurements made at the surface. Geophysical surveys can be useful in the study of most subsurface geologic problems. Geophysics also can contribute to many investigation that are concerned primarily with surface geology. In some investigations, a contribution of drilling and geophysical measurements may provide the optimum costbenefit ratio.

A wide range of physical properties of rocks are used as the bases of particular geophysical methods. The most common in engineering surveys are:

a- elastic properties, used in the seismic methods;
b- electrical properties, used in the electrical and electromagnetic methods;
c- density, in the gravity method; and
d- magnetic properties, used in the magnetic methods.

The choice of a method depends, first, on whether an appropriate physical contrast occurs across the boundary which is being investigated. Criteria, other than geophysical, are the resolution sought, and the relative cost of coverage and time taken, by a particular type of survey; and the nature of the site, for example whether it is water covered, whether it is permissible to explode small charges on it, and so on.

There are many surface and borehole geophysical methods available to investigate and solve subsurface problems. Among the methods commonly employed to solve a variety of problems are:

- Electrical Resistivity
- Seismic Refraction and Reflection
- Ground Penetrating Radar (GPR)
- Electromagnetic Conductivity
- Gravity Survey
- Magnetic Survey

Electrical Resistivity: Electrical resistivity measures the bulk electrical resistance of the subsurface by inducing a current between two electrodes implanted in the subsurface. Electrical resistivity is used to characterize subsurface stratigraphy, water table depth, conductive plumes, and buried wastes. With advances in technology, data collection and inversion methods have been advanced to the point that electrical resistivity is sometimes referred to as ***Electrical Imaging***.

Seismic Reflection and Refraction: Seismic refraction and reflection methods measure the transmission of sound waves through the subsurface generated using a hammer blow or explosive energy source. Individual subsurface layer depths and thickness can be calculated based on the analysis of sonic wave arrival times. Companies use seismic methods to obtain the depth to bedrock surfaces and the water table; to characterize rock type and degree of weathering; and to locate fractures, faults, and buried channels.

Ground Penetrating Radar: Ground Penetrating Radar (GPR) is used to measure changes in the dielectric properties of subsurface materials. Many companies use GPR to delineate subsurface features such as underground storage tanks, buried drums, and utilities. GPR is also used to locate karst-related voids in the subsurface. Other applications include mapping landfilled and excavated areas and site stratigraphy.

Gravity: Gravity measurements detect changes in the earth's gravitational field caused by variations in the density of the soil or rock or engineered structures. Companies use gravity surveys to locate and characterize buried bedrock channels and bedrock structural features. Other applications are to detect voids, caves, and abandoned mines or tunnels.

Magnetic Survey: Magnetic surveying measures the perturbations in the earth's magnetic field caused by changes in concentrations of natural ferrous minerals or by ferrous metals. Companies use magnetic to locate buried ferrous metals such as wastes, drums, or utilities. Other applications include characterizing geologic structures and mineral exploration.

4.2.2 Electrical Methods

4.2.2.1 Basic Theory

Several methods of prospecting for minerals and of investigating geological conditions have been developed using the electrical properties of rocks. The ***resistivity method*** is the one that has been most used in engineering. It depends on the property of resistivity, which is the electrical resistance between opposite faces of a unit cube of a given rock.

The SI unit is the ohm meter (ohm . m). The values of resistivity of rocks and minerals range from 10^{-6} to over 10^{12} ohm m, with no simple relationship of the spread of values to a genetic classification of rocks. Some minerals, like clays, conduct electricity well, but the predominant factor controlling the resistivity of most rocks is the presence or absence of ground water in them, and the relative concentration of electrolyte (dissolved salts). A sand saturated with brackish water will have a very low resistivity. The same sand, if it lay above the water table and had air, not water, in its pores, would have a high resistivity. Crystalline rocks normally have high resistivities, except locally where they are broken by faults or joints and ground water fills the voids.

The electrical conductivity (and its reciprocal) depends mainly on the following factors such as:
1- The volume of voids and fractures within rocks.
2- The continuity and the distribution of the voids (pores) within rocks.
3- The type of liquid in the pores.
4- The degree of porosity and the amount of contained liquids.
5- The degree of saturation.
6- The type of minerals consisting soils and rocks.

In resistivity method, an electric current is introduced into the ground along traverses in the field by means of two non-polarized electrodes (the current electrodes C1 and C2 in Figure 4.1) and the potential difference between two potential electrodes (P1 and P2) is measured over a distance ***a***. The depth of penetration increases with increasing electrode distance (***a***). Thus the simplest method of conducting a resistivity survey is to arrange four electrodes in a straight line on the ground surface. Hence the resistivity value is determined from dividing the measured potential difference by the applied current times the geometric factor (***K***). The geometric factor is calculated from the appropriate distances used between the electrodes . There are mainly two common systems of electrodes configurations, Wenner and Schlumberger configurations.

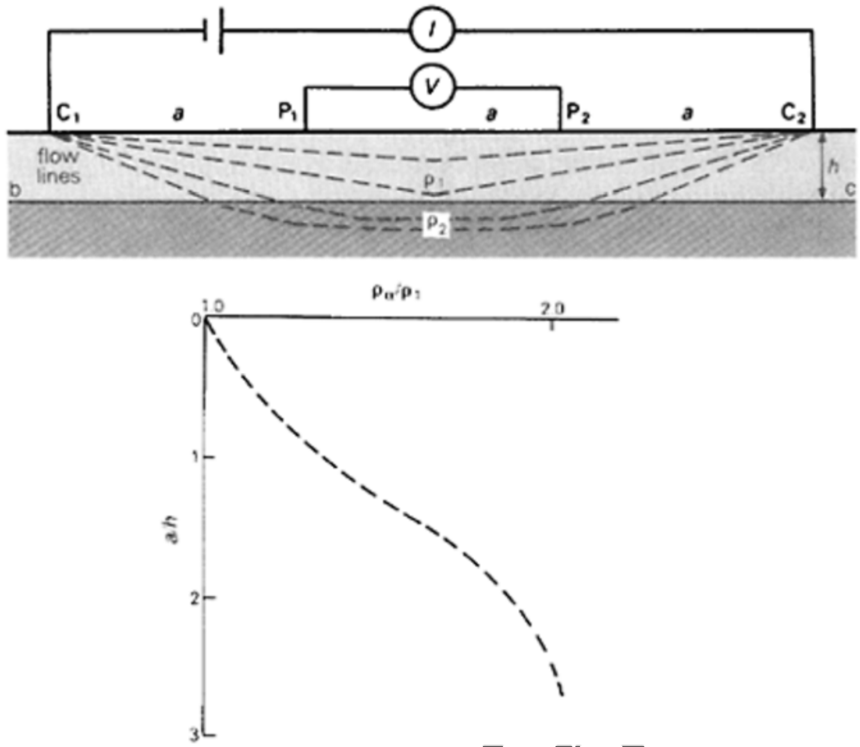


Fig. (4.1). The upper part of the diagram shows the layout of electrodes on the the ground surface using Wenner Array. In the subsurface two layers of different resistivity ρ_1 and ρ_2 ($\rho_1 > \rho_2$) are separated by a horizontal interface bc, and the flux is denser in layer ρ_2 because of its lower resistivity.

4.2.2.2 Types of Arrangements (Configurations)

Wenner Arrangement

In the commonly used ***Wenner arrangement*** (**or *configuration)***, the electrodes are spaced at equal distances C1P1=P1P2=P2C2=***a*** as shown in Figure 4.1. This arrangement is used for the knowledge either the horizontal or the vertical variation depending on the procedure of measure. When the four electrodes are used at a constant spacing a in all measuring points where the electrodes are moved as a group and different profile lines are run across the area. This method is useful for horizontal variations which is called ***electrical profiling*** in which the electrode spacing is kept constant, and the spread is moved laterally in successive steps. But for the knowledge of the vertical variations, the test is repeated after changing the spacing for the same point (i.e. $\boldsymbol{a_1}$ for the first setting and $\mathbf{a_2}$ for the second setting and so on) and again determine the mean resistivity upto the depth equal to the new spacing. This system is useful in studying the vertical changes in the strata with increasing depth at a point which is called ***electrical sounding*** where the centre of the electrode arrangement is kept fixed, but the spacing (***a***) is increased by progressive steps to give deeper and deeper penetration by the current. The shape of the current flowing between current electrodes is homogenous curved paths if the upper layer is homogenous in its thickness and composition and its thickness exceeding the current electrodes spacing. With increasing current electrodes spacing , the current may reach the second layer which has different resistivity, thus the measured resistivity on the ground surface is called the ***apparent resistivity*** **(ρ_a)** which represents the resistivity of both layers. For the Wenner arrangement of electrodes, the resistivity ρ_a of the rock is given by:

$$\rho_a = (V / I) * K_W = R\,K_W \qquad (4.1)$$

where $R = V / I \;;\; K_W = 2\pi a$

$$\rho_a = 2\pi a \,(V / I)$$

$$\rho_a = 2\pi a R \qquad (4.2)$$

where V is the voltage between the potential electrodes, I is the current flowing between the current electrodes, R is the resistance (ohm) and a is the separation of the electrodes.

Where a layer ρ_1 of lower resistivity overlies one of high resistivity ρ_2, the flow lines are distorted from circular arcs, with flow preferentially concentrated in the good conductor ρ_1. If resistivity is measured and calculated from the above equation, it would have a value intermediate between ρ_1 and ρ_2. It is referred to as a value of apparent resistivity ρ_a. The percentage of the total current flowing through the lower rock layer ρ_2 depends on its depth relative to a, and on the contrast in resistivity with ρ_1. As a is increased, the flow of current through layer ρ_2 increases and ρ_a tends towards the value ρ_2. That is, for small values of a, ρ_a approximates to ρ_1 , and for values large compared with the depth to ρ_2 , to ρ_2 .

Figure (4.1) shows in upper part of the diagram the layout of electrodes on the ground surface. In the subsurface two layers of different resistivity ρ_1 and ρ_2 **($\rho_1 > \rho_2$)** are separated by a horizontal interface bc, and the flux is denser in layer ρ_2 because of its lower resistivity. The graph shows how apparent resistivity ρ_a changes with electrode spacing a in a given two-layer case with the interface at depth h, and with a particular contrast between ρ_1 and ρ_2 . Observed results may be interpreted by comparing them with a set of master curves, each of this type and corresponding to a particular contrast of resistivities. Master curves are usually presented on logarithmic scales, and observations are conventionally plotted on log graph paper. If direct current is applied, as shown in this diagram, non-polarizable electrodes must be used. This awkward procedure is obviated by the use of alternating current in most resistivity apparatus.

2- Schlumberger Arrangement

In the commonly used procedure of a ***Schlumberger arrangement*** , the potential electrodes are kept fixed at the center of line **C_1 C_2**, while the current electrodes are moved symmetrically outwards as shown in Figure 4.2.

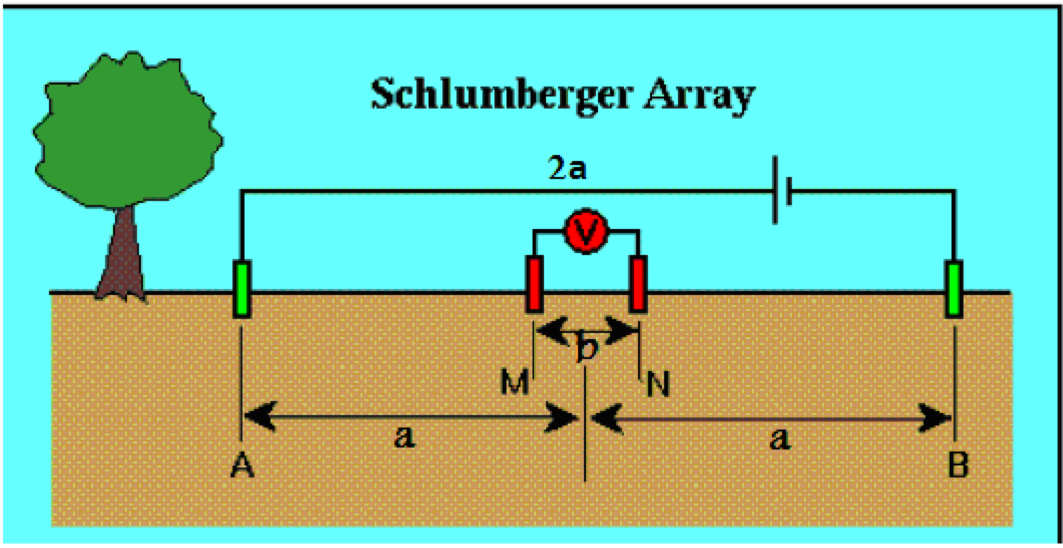


Fig. (4.2). The layout of electrodes using Schlumberger array.

Since only two electrodes are moved, the field procedure with the Schlumberger system is quicker than that with the Wenner system. Also, since the potential electrodes P_1 P_2 are kept fixed, but if at any stage the potential difference between these electrodes becomes too small to be measured accurately then the electrode spacing should be increased and so on.

Then, with using Schlumberger arrangement, the apparent resistivity is calculated by:

$$\rho_a = R\, K_S \qquad (4.3)$$

Where K_S is the geometric factor for Schlumberger arrangement and calculated as follows:

$$K_S = \pi\, [(AB)^2 - (MN)^2] / 4(MN) \;\; ; \; or \quad K_S = [\, \pi(AM) * (AN)\,] / (MN) \qquad (4.4)$$

$$\rho_a = \pi\, R\, [(AB)^2 - (MN)^2] / 4(MN \;\; ; \; or \quad \rho_a = R\, [\, \pi(AM) * (AN)\,] / (MN) \qquad (4.5)$$

4.2.2.3 Applications of Resistivity Methods

1- Detecting the depth of bed rock which is very important for major engineering structures such as dams , reservoir and tunnels.
2- Detecting the subsurface geologic structures, such as faults orientation and seepage directions.
3- Detecting the locations position and type of building materials, such as gravels, sands and clays.

4- Studying the groundwater conditions by identifying their water levels and the type of aquifers.
5- Detecting the lateral and vertical variations in rock layers from the variation in the type of sediments.

EXAMPLE 4.1: An electrical survey was carried out using Wenner arrangement to detect groundwater table. The used spacing was (a = 2 m), the current was (I = 60 mA) , and the potential difference was (V = 240 mV). Calculate the apparent resistivity of the surface layer?

$R = V / I$ = 240 mV / 60 mA = 4.0 mohm
$\rho_a = 2\pi\, a\, R$
$\rho_a = 2\pi$ * (2 m) (4 mohm)
ρ_a = 50.0 mohm.m

EXAMPLE 4.2: An electrical survey was made using Schlumberger configuration in a station, the measurements were tabulated as follows ; Find:
1- the geometric factor (K_S) ; 2- the apparent resistivity

AB / 2 (m)	*MN / 2 (m)*	*R (ohm)*
1.5	0.5	2.637
2.0	0.5	1.297
3.0	0.5	0.4865
4.0	0.5	0.2188
5.0	0.5	0.131
6.0	1.0	0.168
8.0	1.0	0.085
10.0	1.0	0.0532

AB / 2 (m)	*MN / 2 (m)*	*R (ohm)*	K_S	ρ_a *(ohm.m)*
1.5	0.5	2.637	6.28	16.56
2.0	0.5	1.297	11.78	14.69
3.0	0.5	0.4865	27.49	13.37
4.0	0.5	0.2188	49.48	10.82
5.0	0.5	0.131	77.75	10.19
6.0	1.0	0.168	54.48	9.15
8.0	1.0	0.085	98.96	8.41
10.0	1.0	0.0532	155.51	8.27

EXAMPLE 4.3: For the above example (4.2) ; it is required to find:
1- the geometric factor (K_S) using the formula $K_S = [\,\pi(AM) * (AN)\,] / (MN)$
2- the apparent resistivity.
3- compare both results .

AB / 2 (m)	*MN / 2 (m)*	*R (ohm)*	K_S	ρ_a *(ohm.m)*
1.5	0.5	2.637		
2.0	0.5	1.297		
3.0	0.5	0.4865		
4.0	0.5	0.2188		
5.0	0.5	0.131		
6.0	1.0	0.168		
8.0	1.0	0.085		
10.0	1.0	0.0532		

4.2.3 Seismic Method

4.2.3.1 Introduction

The applied seismic methods are used to determine the distribution of elastic properties under a site, as a step to inferring what rocks and structures are present. The behavior of wave motions in rocks is dependent on the elastic properties of the rocks. Two parameters are required to describe the elasticity of an isotropic material. Young's Modulus and the Poisson's ratio are the pair most familiar to engineers. The two aspects of elasticity, ***resistance to change of volume*** and ***resistance to change of shape***, are, however, most simply described by the ***coefficient of incompressibility*** (***bulk modulus***) ***k*** and the ***coefficient of rigidity*** (***shear modulus***) ***μ***.

4.2.3.2 Types of Waves

There are two types of elastic wave

1- **Primary (or Longitudinal)waves (P-Waves) :** In ***longitudinal waves***, the particle motion is parallel to the direction of propagation, and each part of the rock is periodically compressed and dilated by the wave motion. Their velocity of propagation, V_P , is given by:

$$V_P = [(k + 4/3\ G) / \rho\,]^{1/2} \qquad (4.6)$$

where k is the bulk modulus, G the shear modulus and ρ is the density of the rocks, or

$$\rho\, V^2_P = E\,(1 - \mu) / [(1+ \mu)(1 - 2\mu)] \qquad (4.7)$$