

## Introduction to Soil Mechanics

The term "soil" can have different meanings, depending upon the field in which it is considered. To a geologist, it is the material in the relative thin zone of the Earth's surface within which roots occur, and which are formed as the products of past surface processes. The rest of the crust is grouped under the term "rock".

To a pedologist, it is the substance existing on the surface, which supports plant life. To an engineer, it is a material that can be: built on: foundations of buildings, bridges built in: basements, culverts, tunnels built with: embankments, roads, dams supported: retaining walls

Soil Mechanics is a discipline of Civil Engineering involving the study of soil, its behaviour and application as an engineering material. Soil Mechanics is the application of laws of mechanics and hydraulics to engineering problems dealing with sediments and other unconsolidated accumulations of solid particles, which are produced by the mechanical and chemical disintegration of rocks, regardless of whether or not they contain an admixture of organic constituents.

Soil consists of a multiphase aggregation of solid particles, water, and air. This fundamental composition gives rise to unique engineering properties, and the description of its mechanical behavior requires some of the most classic principles of engineering mechanics.

Engineers are concerned with soil's mechanical properties: permeability, stiffness, and strength. These depend primarily on the nature of the soil grains, the current stress, the water content and unit weight.

**Formation of Soils:** In the Earth's surface, rocks extend upto as much as 20 km depth. The major rock types are categorized as igneous, sedimentary, and metamorphic.

**Igneous rocks:** formed from crystalline bodies of cooled magma.

**Sedimentary rocks:** formed from layers of cemented sediments.

**Metamorphic rocks:** formed by the alteration of existing rocks due to heat from igneous intrusions or pressure due to crustal movement.

Soils are formed from materials that have resulted from the disintegration of rocks by various processes of physical and chemical weathering. The nature and structure of a given soil depends on the processes and conditions that formed it:

Breakdown of parent rock: weathering, decomposition, erosion.

Transportation to site of final deposition: gravity, flowing water, ice, wind.

Environment of final deposition: flood plain, river terrace, glacial moraine, lacustrine or marine.

Subsequent conditions of loading and drainage: little or no surcharge, heavy surcharge due to ice or overlying deposits, change from saline to freshwater, leaching, contamination.

All soils originate, directly or indirectly, from different rock types.

Physical weathering reduces the size of the parent rock material, without any change in the original composition of the parent rock. Physical or mechanical processes taking place on the earth's surface include the actions of water, frost, temperature changes, wind and ice. They cause disintegration and the products are mainly coarse soils.

The main processes involved are exfoliation, unloading, erosion, freezing, and thawing. The principal cause is climatic change. In exfoliation, the outer shell separates from the main rock. Heavy rain and wind cause erosion of the rock surface. Adverse temperature changes produce fragments due to different thermal coefficients of rock minerals. The effect is more for freeze-thaw cycles.

Chemical weathering not only breaks up the material into smaller particles but alters the nature of the original parent rock itself. The main processes responsible are hydration, oxidation, and carbonation. New compounds are formed due to the chemical alterations.

Rain water that comes in contact with the rock surface reacts to form hydrated oxides, carbonates and sulphates. If there is a volume increase, the disintegration continues. Due to leaching, water-soluble materials are washed away and rocks lose their cementing properties.

Chemical weathering occurs in wet and warm conditions and consists of degradation by decomposition and/or alteration. The results of chemical

weathering are generally fine soils with altered mineral grains.

The effects of weathering and transportation mainly determine the basic nature of the soil (size, shape, composition and distribution of the particles).

The environment into which deposition takes place, and the subsequent geological events that take place there, determine the state of the soil (density, moisture content) and the structure or fabric of the soil (bedding, stratification, occurrence of joints or fissures)

Transportation agencies can be combinations of gravity, flowing water or air, and moving ice. In water or air, the grains become sub-rounded or rounded, and the grain sizes get sorted so as to form poorly-graded deposits. In moving ice, grinding and crushing occur, size distribution becomes wider forming well-graded deposits.

In running water, soil can be transported in the form of suspended particles, or by rolling and sliding along the bottom. Coarser particles settle when a decrease in velocity occurs, whereas finer particles are deposited further downstream. In still water, horizontal layers of successive sediments are formed, which may change with time, even seasonally or daily.

Wind can erode, transport and deposit fine-grained soils. Wind-blown soil is generally uniformly-graded.

A glacier moves slowly but scours the bedrock surface over which it passes. Gravity transports materials along slopes without causing much alteration.

**Soil Types:** Soils as they are found in different regions can be classified into two broad categories: 1) Residual soils (2) Transported soils

**Residual Soils:** Residual soils are found at the same location where they have been formed. Generally, the depth of residual soils varies from 5 to 20 m.

Chemical weathering rate is greater in warm, humid regions than in cold, dry regions causing a faster breakdown of rocks. Accumulation of residual soils takes place as the rate of rock decomposition exceeds the rate of erosion or transportation of the weathered material. In humid regions, the presence of surface vegetation reduces the possibility of soil transportation.

As leaching action due to percolating surface water decreases with depth, there is a corresponding decrease in the degree of chemical weathering from the ground surface downwards. This results in a gradual reduction of residual soil formation with depth, until unaltered rock is found.

Residual soils comprise of a wide range of particle sizes, shapes and composition.

**Transported Soils:** Weathered rock materials can be moved from their original site to new locations by one or more of the transportation agencies to form transported soils. Transported soils are classified based on the mode of transportation and the final deposition environment.

(a) Soils that are carried and deposited by rivers are called alluvial deposits. (b) Soils that are deposited by flowing water or surface runoff while entering a lake are called lacustrine deposits. Alternate layers are formed in different seasons depending on flow rate.

(c) If the deposits are made by rivers in sea water, they are called marine deposits. Marine deposits contain both particulate material brought from the shore as well as organic remnants of marine life forms.

(d) Melting of a glacier causes the deposition of all the materials scoured by it leading to formation of glacial deposits.

(e) Soil particles carried by wind and subsequently deposited are known as aeolian deposits.

**Phase Relations of Soils:** Soil is not a coherent solid material like steel and concrete, but is a particulate material. Soils, as they exist in nature, consist of solid particles (mineral grains, rock fragments) with water and air in the voids between the particles. The water and air contents are readily changed by changes in ambient conditions and location.

As the relative proportions of the three phases vary in any soil deposit, it is useful to consider a soil model which will represent these phases distinctly and properly quantify the amount of each phase. A schematic diagram of the three-phase system is shown in terms of weight and volume symbols respectively for soil solids, water, and air. The weight of air can be

neglected.

The soil model is given dimensional values for the solid, water and air components. Total volume,  $V = V_s + V_w + V_v$

**Three-phase System:** Soils can be partially saturated (with both air and water present), or be fully saturated (no air content) or be perfectly dry (no water content).

In a saturated soil or a dry soil, the three-phase system thus reduces to two phases only, as shown.

For the purpose of engineering analysis and design, it is necessary to express relations between the weights and the volumes of the three phases.

The various relations can be grouped into:

Volume relations      Weight relations      Inter-relations      Volume Relations

As the amounts of both water and air are variable, the volume of solids is taken as the reference quantity. Thus, several relational volumetric quantities may be defined. The following are the basic volume relations:

1. Void ratio ( $e$ ) is the ratio of the volume of voids ( $V_v$ ) to the volume of soil solids ( $V_s$ ), and is expressed as a decimal.

2. Porosity ( $n$ ) is the ratio of the volume of voids to the total volume of soil ( $V$ ), and is expressed as a percentage.

Void ratio and porosity are inter-related to each other as follows:  
and

3. The volume of water ( $V_w$ ) in a soil can vary between zero (i.e. a dry soil) and the volume of voids. This can be expressed as the degree of saturation ( $S$ ) in percentage.

For a dry soil,  $S = 0\%$ , and for a fully saturated soil,  $S = 100\%$ .

4. Air content ( $ac$ ) is the ratio of the volume of air ( $V_a$ ) to the volume of voids.

5. Percentage air voids ( $na$ ) is the ratio of the volume of air to the total volume.

**Weight Relations:** Density is a measure of the quantity of mass in a unit volume of material. Unit weight is a measure of the weight of a unit volume of material. Both can be used interchangeably. The units of density are  $\text{ton/m}^3$ ,  $\text{kg/m}^3$  or  $\text{g/cm}^3$ . The following are the basic weight relations:

1. The ratio of the mass of water present to the mass of solid particles is called the water content ( $w$ ), or sometimes the moisture content.

Its value is 0% for dry soil and its magnitude can exceed 100%.

2. The mass of solid particles is usually expressed in terms of their particle unit weight or specific gravity ( $G_s$ ) of the soil grain solids. where  $\gamma_s = \text{Unit weight of water}$

For most inorganic soils, the value of  $G_s$  lies between 2.60 and 2.80. The presence of organic material reduces the value of  $G_s$ .

3. Dry unit weight is a measure of the amount of solid particles per unit volume.

4. Bulk unit weight is a measure of the amount of solid particles plus water per unit volume.

5. Saturated unit weight is equal to the bulk density when the total voids is filled up with water.

6. Buoyant unit weight or submerged unit weight is the effective mass per unit volume when the soil is submerged below standing water or below the ground water table.

**Inter-Relations:** It is important to quantify the state of a soil immediately after receiving in the laboratory and prior to commencing other tests. The water content and unit weight are particularly important, since they may change during transportation and storage.

Some physical state properties are calculated following the practical measurement of others. For example, dry unit weight can be determined from bulk unit weight and water content. The following are some inter-relations:

## Worked Examples

### Soil Classification

It is necessary to adopt a formal system of soil description and classification in order to describe the various materials found in ground investigation. Such a system must be meaningful and concise in an engineering context, so that engineers will be able to understand and interpret.

It is important to distinguish between description and classification:

Description of soil is a statement that describes the physical nature and state of the soil. It can be a description of a sample, or a soil in situ. It is arrived at by using visual examination, simple tests, observation of site conditions, geological history, etc.

Classification of soil is the separation of soil into classes or groups each having similar characteristics and potentially similar behaviour. A classification for engineering purposes should be based mainly on mechanical properties: permeability, stiffness, strength. The class to which a soil belongs can be used in its description.

The aim of a classification system is to establish a set of conditions which will allow useful comparisons to be made between different soils. The system must be simple. The relevant criteria for classifying soils are the size distribution of particles and the plasticity of the soil.

**Particle Size Distribution:** For measuring the distribution of particle sizes in a soil sample, it is necessary to conduct different particle-size tests.

Wet sieving is carried out for separating fine grains from coarse grains by washing the soil specimen on a 75 micron sieve mesh.

Dry sieve analysis is carried out on particles coarser than 75 micron. Samples (with fines removed) are dried and shaken through a set of sieves of descending size. The weight retained in each sieve is measured. The cumulative percentage quantities finer than the sieve sizes (passing each given sieve size) are then determined.

The resulting data is presented as a distribution curve with grain size along x-axis (log scale) and percentage passing along y-axis (arithmetic scale).

Sedimentation analysis is used only for the soil fraction finer than 75 microns. Soil particles are allowed to settle from a suspension. The decreasing density of the suspension is measured at various time intervals. The procedure is based on the principle that in a suspension, the terminal velocity of a spherical particle is governed by the diameter of the particle and the properties of the suspension.

In this method, the soil is placed as a suspension in a jar filled with distilled water to which a deflocculating agent is added. The soil particles are then allowed to settle down. The concentration of particles remaining in the suspension at a particular level can be determined by using a hydrometer. Specific gravity readings of the solution at that same level at different time intervals provide information about the size of particles that have settled down and the mass of soil remaining in solution.

The results are then plotted between % finer (passing) and log size.

**Grain-Size Distribution Curve:** The size distribution curves, as obtained from coarse and fine grained portions, can be combined to form one complete grain-size distribution curve (also known as grading curve). A typical grading curve is shown.

From the complete grain-size distribution curve, useful information can be obtained such as:

1. Grading characteristics, which indicate the uniformity and range in grain-size distribution.
2. Percentages (or fractions) of gravel, sand, silt and clay-size.

**Grading Characteristics:** A grading curve is a useful aid to soil description. The geometric properties of a grading curve are called grading characteristics.

To obtain the grading characteristics, three points are located first on the grading curve.  $D_{60}$  = size at 60% finer by weight ;  $D_{30}$  = size at 30% finer by weight  
 $D_{10}$  = size at 10% finer by weight

The grading characteristics are then determined as follows:

1. Effective size =  $D_{10}$ ; 2. Uniformity coefficient, 3. Curvature coefficient, Both  $C_u$  and  $C_c$  will be 1 for a single-sized soil.  $C_u > 5$  indicates a well-graded soil, i.e. a soil which has a distribution of particles over a wide size range.  $C_c$  between 1 and 3 also indicates a well-graded soil.  $C_u < 3$  indicates a uniform soil, i.e. a soil which has a very narrow particle size range.

**Consistency of Soils:** The consistency of a fine-grained soil refers to its firmness, and it varies with the water content of the soil. A gradual increase in water content causes the soil to change from solid to semi-solid to plastic to liquid states. The water contents at which the consistency changes from one state to the other are called consistency limits (or Atterberg limits).

The three limits are known as the shrinkage limit (WS), plastic limit (WP), and liquid limit (WL) as shown. The values of these limits can be obtained from laboratory tests.

Two of these are utilised in the classification of fine soils:

Liquid limit (WL) - change of consistency from plastic to liquid state

Plastic limit (WP) - change of consistency from brittle/crumbly to plastic state

The difference between the liquid limit and the plastic limit is known as the plasticity index (IP), and it is in this range of water content that the soil has a plastic consistency. The consistency of most soils in the field will be plastic or semi-solid.

#### Indian Standard Soil Classification System

**Classification Based on Grain Size:** The range of particle sizes encountered in soils is very large: from boulders with dimension of over 300 mm down to clay particles that are less than 0.002 mm. Some clays contain particles less than 0.001 mm in size which behave as colloids, i.e. do not settle in water.

In the Indian Standard Soil Classification System (ISSCS), soils are classified into groups according to size, and the groups are further divided into coarse, medium and fine sub-groups.

The grain-size range is used as the basis for grouping soil particles into boulder, cobble, gravel, sand, silt or clay.

Very coarse soils Boulder size > 300 mm Cobble size 80 - 300 mm Coarse soils Gravel size (G) Coarse 20 - 80 mm Fine 4.75 - 20 mm Sand size (S) Coarse 2 - 4.75 mm

Medium 0.425 - 2 mm Fine 0.075 - 0.425 mm Fine soils Silt size (M) 0.002 - 0.075 mm Clay size (C) < 0.002 mm Gravel, sand, silt, and clay are represented by group symbols G, S, M, and C respectively.

Physical weathering produces very coarse and coarse soils. Chemical weathering produce generally fine soils.

Coarse-grained soils are those for which more than 50% of the soil material by weight has particle sizes greater than 0.075 mm. They are basically divided into either gravels (G) or sands (S).

According to gradation, they are further grouped as well-graded (W) or poorly graded (P). If fine soils are present, they are grouped as containing silt fines (M) or as containing clay fines (C).

For example, the combined symbol SW refers to well-graded sand with no fines.

Both the position and the shape of the grading curve for a soil can aid in establishing its identity and description. Some typical grading curves are shown.

Curve A - a poorly-graded medium SAND Curve B - a well-graded GRAVEL-SAND (i.e. having equal amounts of gravel and sand) Curve C - a gap-graded COBBLES-SAND Curve D - a sandy SILT Curve E - a silty CLAY (i.e. having little amount of sand)

Fine-grained soils are those for which more than 50% of the material has particle sizes less than 0.075 mm. Clay particles have a flaky shape to which water adheres, thus imparting the property of plasticity.

A plasticity chart, based on the values of liquid limit (WL) and

plasticity index (IP), is provided in ISSCS to aid classification. The 'A' line in this chart is expressed as  $IP = 0.73 (WL - 20)$ .

Depending on the point in the chart, fine soils are divided into clays (C), silts (M), or organic soils (O). The organic content is expressed as a percentage of the mass of organic matter in a given mass of soil to the mass of the dry soil solids. Three divisions of plasticity are also defined as follows.  
Low plasticity  $WL < 35\%$                       Intermediate plasticity  $35\% < WL < 50\%$                       High plasticity  $WL > 50\%$

The 'A' line and vertical lines at WL equal to 35% and 50% separate the soils into various classes. For example, the combined symbol CH refers to clay of high plasticity.

Soil classification using group symbols is as follows:      Group Symbol  
Classification

Coarse soils GW: Well-graded GRAVEL GP ; Poorly-graded GRAVEL GM; Silty GRAVEL  
GC: Clayey GRAVEL SW: Well-graded SAND SP: poorly-graded SAND SM : Silty  
SAND SC: Clayey SAND Fine soils ML SILT of low plasticity MI: SILT of  
intermediate plasticity MH: SILT of high plasticity CL CLAY of low  
plasticity: CI CLAY of intermediate plasticity CH; CLAY of high plasticity OL  
Organic soil of low plasticity OI: organic soil of intermediate plasticity OH:  
organic soil of high plasticity Pt Peat

Activity: "Clayey soils" necessarily do not consist of 100% clay size particles. The proportion of clay mineral flakes ( $< 0.002$  mm size) in a fine soil increases its tendency to swell and shrink with changes in water content. This is called the activity of the clayey soil, and it represents the degree of plasticity related to the clay content.                      Activity = (Plasticity index) / (% clay particles by weight)

Classification as per activity is:                      Activity      Classification <  
0.75 Inactive      0.75 - 1.25 Normal      > 1.25 Active

Liquidity Index: In fine soils, especially with clay size content, the existing state is dependent on the current water content (w) with respect to the consistency limits (or Atterberg limits). The liquidity index (LI) provides a quantitative measure of the present state.

Classification as per liquidity index is:

Liquidity index      Classification      > 1 Liquid      0.75 - 1.00 Very soft      0.50 - 0.75  
Soft      0.25 - 0.50 Medium stiff      0 - 0.25 Stiff      < 0 Semi-solid

Visual Classification: Soils possess a number of physical characteristics which can be used as aids to identification in the field. A handful of soil rubbed through the fingers can yield the following:

SAND (and coarser) particles are visible to the naked eye.

SILT particles become dusty when dry and are easily brushed off hands.

CLAY particles are sticky when wet and hard when dry, and have to be scraped or washed off hands.

Worked Example

Formation of Clay Minerals: A soil particle may be a mineral or a rock fragment. A mineral is a chemical compound formed in nature during a geological process, whereas a rock fragment has a combination of one or more minerals. Based on the nature of atoms, minerals are classified as silicates, aluminates, oxides, carbonates and phosphates. Out of these, silicate minerals are the most important as they influence the properties of clay soils. Different arrangements of atoms in the silicate minerals give rise to different silicate structures.

Basic Structural Units: Soil minerals are formed from two basic structural units: tetrahedral and octahedral. Considering the valencies of the atoms forming the units, it is clear that the units are not electrically neutral and as such do not exist as single units.

The basic units combine to form sheets in which the oxygen or hydroxyl ions are shared among adjacent units. Three types of sheets are thus formed, namely silica sheet, gibbsite sheet and brucite sheet.

Isomorphous substitution is the replacement of the central atom of the tetrahedral or octahedral unit by another atom during the formation of the sheets. The sheets then combine to form various two-layer or three-layer sheet minerals. As the basic units of clay minerals are sheet-like structures, the particle formed from stacking of the basic units is also plate-like. As a result, the surface area per unit mass becomes very large.

**Structure of Clay Minerals:** A tetrahedral unit consists of a central silicon atom that is surrounded by four oxygen atoms located at the corners of a tetrahedron. A combination of tetrahedrons forms a silica sheet.

An octahedral unit consists of a central ion, either aluminium or magnesium, that is surrounded by six hydroxyl ions located at the corners of an octahedron. A combination of aluminium-hydroxyl octahedrons forms a gibbsite sheet, whereas a combination of magnesium-hydroxyl octahedrons forms a brucite sheet.

**Two-layer Sheet Minerals:** Kaolinite and halloysite clay minerals are the most common.

**Kaolinite Mineral** The basic kaolinite unit is a two-layer unit that is formed by stacking a gibbsite sheet on a silica sheet. These basic units are then stacked one on top of the other to form a lattice of the mineral. The units are held together by hydrogen bonds. The strong bonding does not permit water to enter the lattice. Thus, kaolinite minerals are stable and do not expand under saturation. Kaolinite is the most abundant constituent of residual clay deposits.

**Halloysite Mineral :** The basic unit is also a two-layer sheet similar to that of kaolinite except for the presence of water between the sheets.

**Three-layer Sheet Minerals:** Montmorillonite and illite clay minerals are the most common. A basic three-layer sheet unit is formed by keeping one silica sheet each on the top and at the bottom of a gibbsite sheet. These units are stacked to form a lattice as shown.

**Montmorillonite Mineral :** The bonding between the three-layer units is by van der Waals forces. This bonding is very weak and water can enter easily. Thus, this mineral can imbibe a large quantity of water causing swelling. During dry weather, there will be shrinkage.

**Illite Mineral :** Illite consists of the basic montmorillonite units but are bonded by secondary valence forces and potassium ions, as shown. There is about 20% replacement of aluminium with silicon in the gibbsite sheet due to isomorphous substitution. This mineral is very stable and does not swell or shrink.

**Fine Soil Fabric:** Natural soils are rarely the same from one point in the ground to another. The content and nature of grains varies, but more importantly, so does the arrangement of these. The arrangement and organisation of particles and other features within a soil mass is termed its fabric.

CLAY particles are flaky. Their thickness is very small relative to their length & breadth, in some cases as thin as 1/100th of the length. They therefore have high specific surface values. These surfaces carry negative electrical charge, which attracts positive ions present in the pore water. Thus a lot of water may be held as adsorbed water within a clay mass.

**Stresses in the Ground:**

**Total Stress:** When a load is applied to soil, it is carried by the solid grains and the water in the pores. The total vertical stress acting at a point below the ground surface is due to the weight of everything that lies above, including soil, water, and surface loading. Total stress thus increases with depth and with unit weight. Vertical total stress at depth  $z$ ,  $\sigma_v = \gamma_s \cdot z$

Below a water body, the total stress is the sum of the weight of the soil up to the surface and the weight of water above this.  $\sigma_v = \gamma_s \cdot z + \gamma_w \cdot W$

The total stress may also be denoted by  $\sigma_z$  or just  $\sigma$ . It varies with changes in water level and with excavation.

**Pore Water Pressure:** The pressure of water in the pores of the soil is called pore water pressure ( $u$ ). The magnitude of pore water pressure depends on:

the depth below the water table. the conditions of seepage flow.

Under hydrostatic conditions, no water flow takes place, and the pore pressure at a given point is given by

$u = \gamma_w \cdot h$  where  $h$  = depth below water table or overlying water surface

It is convenient to think of pore water pressure as the pressure exerted by a column of water in an imaginary standpipe inserted at the given point.

The natural level of ground water is called the water table or the phreatic surface. Under conditions of no seepage flow, the water table is horizontal. The magnitude of the pore water pressure at the water table is zero. Below the water table, pore water pressures are positive.

**Principle of Effective Stress:** The principle of effective stress was enunciated by Karl Terzaghi in the year 1936. This principle is valid only for saturated soils, and consists of two parts:

1. At any point in a soil mass, the effective stress (represented by  $\sigma'$ ) is related to total stress ( $\sigma$ ) and pore water pressure ( $u$ ) as

$$\sigma' = \sigma - u$$

Both the total stress and pore water pressure can be measured at any point.

2. All measurable effects of a change of stress, such as compression and a change of shearing resistance, are exclusively due to changes in effective stress.

Compression =

Shear Strength =

In a saturated soil system, as the voids are completely filled with water, the pore water pressure acts equally in all directions.

The effective stress is not the exact contact stress between particles but the distribution of load carried by the soil particles over the area considered. It cannot be measured and can only be computed.

If the total stress is increased due to additional load applied to the soil, the pore water pressure initially increases to counteract the additional stress. This increase in pressure within the pores might cause water to drain out of the soil mass, and the load is transferred to the solid grains. This will lead to the increase of effective stress.

**Effective Stress in Unsaturated Zone:** Above the water table, when the soil is saturated, pore pressure will be negative (less than atmospheric). The height above the water table to which the soil is saturated is called the capillary rise, and this depends on the grain size and the size of pores. In coarse soils, the capillary rise is very small.

Between the top of the saturated zone and the ground surface, the soil is partially saturated, with a consequent reduction in unit weight. The pore pressure in a partially saturated soil consists of two components: Pore water pressure =  $u_w$  Pore air pressure =  $u_a$

Water is incompressible, whereas air is compressible. The combined effect is a complex relationship involving partial pressures and the degree of saturation of the soil.

**Effective Stress Under Hydrodynamic Conditions**

There is a change in pore water pressure in conditions of seepage flow within the ground. Consider seepage occurring between two points P and Q. The potential driving the water flow is the hydraulic gradient between the two points, which is equal to the head drop per unit length. In steady state seepage, the gradient remains constant.

Hydraulic gradient from P to Q,  $i = \frac{h}{L}$

As water percolates through soil, it exerts a drag on soil particles it comes in contact with. Depending on the flow direction, either downward or upward, the drag either increases or decreases inter-particle contact forces.

A downward flow increases effective stress. In contrast, an upward flow opposes the force of gravity and can even cause to counteract completely the contact forces. In such a situation, effective stress is reduced to zero and the soil behaves like a very viscous liquid. Such a state is known as quick sand condition. In nature, this condition is usually observed in coarse silt or fine sand subject to artesian conditions.

At the bottom of the soil column,

During quick sand condition, the effective stress is reduced to zero.

where  $i_{cr}$  = critical hydraulic gradient



This shows that when water flows upward under a hydraulic gradient of about 1, it completely neutralizes the force on account of the weight of particles, and thus leaves the particles suspended in water.

**The Importance of Effective Stress:** At any point within the soil mass, the magnitudes of both total stress and pore water pressure are dependent on the ground water position. With a shift in the water table due to seasonal fluctuations, there is a resulting change in the distribution in pore water pressure with depth.

Changes in water level below ground result in changes in effective stresses below the water table. A rise increases the pore water pressure at all elevations thus causing a decrease in effective stress. In contrast, a fall in the water table produces an increase in the effective stress.

Changes in water level above ground do not cause changes in effective stresses in the ground below. A rise above ground surface increases both the total stress and the pore water pressure by the same amount, and consequently effective stress is not altered.

In some analyses it is better to work with the changes of quantity, rather than in absolute quantities. The effective stress expression then becomes:  $\sigma' = \sigma - u$

If both total stress and pore water pressure change by the same amount, the effective stress remains constant.

Total and effective stresses must be distinguishable in all calculations. Ground movements and instabilities can be caused by changes in total stress, such as caused by loading by foundations and unloading due to excavations. They can also be caused by changes in pore water pressures, such as failure of slopes after rainfall.

**Worked Examples**

**Permeability of Soils:** pressure, Elevation and Total Heads

In soils, the interconnected pores provide passage for water. A large number of such flow paths act together, and the average rate of flow is termed the coefficient of permeability, or just permeability. It is a measure of the ease that the soil provides to the flow of water through its pores.

At point A, the pore water pressure ( $u$ ) can be measured from the height of water in a standpipe located at that point.

The height of the water column is the pressure head ( $h_w$ ).  $h_w = u / \gamma_w$

To identify any difference in pore water pressure at different points, it is necessary to eliminate the effect of the points of measurement. With this in view, a datum is required from which locations are measured.

The elevation head ( $h_z$ ) of any point is its height above the datum line. The height of water level in the standpipe above the datum is the piezometric head ( $h$ ).

$$h = h_z + h_w$$

Total head consists of three components: elevation head, pressure head, and velocity head. As seepage velocity in soils is normally low, velocity head is ignored, and total head becomes equal to the piezometric head. Due to the low seepage velocity and small size of pores, the flow of water in the pores is steady and laminar in most cases. Water flow takes place between two points in soil due to the difference in total heads.

**Darcy's Law:** Darcy's law states that there is a linear relationship between flow velocity ( $v$ ) and hydraulic gradient ( $i$ ) for any given saturated soil under steady laminar flow conditions.

If the rate of flow is  $q$  (volume/time) through cross-sectional area ( $A$ ) of the soil mass, Darcy's Law can be expressed as  $v = q/A = k \cdot i$

where  $k$  = permeability of the soil

$i = \Delta h / L$        $\Delta h$  = difference in total heads       $L$  = length of the soil mass

The flow velocity ( $v$ ) is also called the Darcian velocity or the superficial velocity. It is different from the actual velocity inside the soil pores, which is known as the seepage velocity,  $v_s$ . At the particulate level, the water

follows a tortuous path through the pores. Seepage velocity is always greater than the superficial velocity, and it is expressed as:

where  $AV$  = Area of voids on a cross section normal to the direction of flow  
 $n$  = porosity of the soil

#### Permeability of Different Soils

Permeability ( $k$ ) is an engineering property of soils and is a function of the soil type. Its value depends on the average size of the pores and is related to the distribution of particle sizes, particle shape and soil structure. The ratio of permeabilities of typical sands/gravels to those of typical clays is of the order of 106. A small proportion of fine material in a coarse-grained soil can lead to a significant reduction in permeability. For different soil types as per grain size, the orders of magnitude for permeability are as follows:

Soil  $k$  (cm/sec) Gravel 100 Coarse sand 100 to 10<sup>-1</sup> Medium sand 10<sup>-1</sup> to 10<sup>-2</sup> Fine sand 10<sup>-2</sup> to 10<sup>-3</sup> Silty sand 10<sup>-3</sup> to 10<sup>-4</sup> Silt 1 x 10<sup>-5</sup>  
Clay 10<sup>-7</sup> to 10<sup>-9</sup>

**Factors affecting Permeability:** In soils, the permeant or pore fluid is mostly water whose variation in property is generally very less. Permeability of all soils is strongly influenced by the density of packing of the soil particles, which can be represented by void ratio ( $e$ ) or porosity ( $n$ ).

**For Sands:** In sands, permeability can be empirically related to the square of some representative grain size from its grain-size distribution. For filter sands, Allen Hazen in 1911 found that  $k = 100 (D_{10})^2$  cm/s where  $D_{10}$  = effective grain size in cm.

Different relationships have been attempted relating void ratio and permeability, such as  $k = \frac{e^3}{1+e}$ , and  $k = \frac{e^2}{1+e}$ . They have been obtained from the Kozeny-Carman equation for laminar flow in saturated soils.

where  $k_0$  and  $k_T$  are factors depending on the shape and tortuosity of the pores respectively,  $SS$  is the surface area of the solid particles per unit volume of solid material, and  $\gamma_w$  and  $\mu$  are unit weight and viscosity of the pore water. The equation can be reduced to a simpler form as

**For Silts and Clays:** For silts and clays, the Kozeny-Carman equation does not work well, and  $\log k$  versus  $e$  plot has been found to indicate a linear relationship.

For clays, it is typically found that

where  $C_k$  is the permeability change index and  $e_k$  is a reference void ratio.

#### Laboratory Measurement of Permeability

**Constant Head Flow:** Constant head permeameter is recommended for coarse-grained soils only since for such soils, flow rate is measurable with adequate precision. As water flows through a sample of cross-section area  $A$ , steady total head drop  $h$  is measured across length  $L$ .

Permeability  $k$  is obtained from: **Falling Head Flow**

Falling head permeameter is recommended for fine-grained soils.

Total head  $h$  in standpipe of area  $a$  is allowed to fall. Hydraulic gradient varies with time. Heads  $h_1$  and  $h_2$  are measured at times  $t_1$  and  $t_2$ . At any time  $t$ , flow through the soil sample of cross-sectional area  $A$  is

----- (1)

Flow in unit time through the standpipe of cross-sectional area  $a$  is

= ----- (2)

Equating (1) and (2), or Integrating between the limits,

**Field Tests for Permeability**

Field or in-situ measurement of permeability avoids the difficulties involved in obtaining and setting up undisturbed samples in a permeameter. It also provides information about bulk permeability, rather than merely the permeability of a small sample.

A field permeability test consists of pumping out water from a main well

and observing the resulting drawdown surface of the original horizontal water table from at least two observation wells. When a steady state of flow is reached, the flow quantity and the levels in the observation wells are noted.

Two important field tests for determining permeability are: Unconfined flow pumping test, and confined flow pumping test.

**Unconfined Flow Pumping Test:** In this test, the pumping causes a drawdown in an unconfined (i.e. open surface) soil stratum, and generates a radial flow of water towards the pumping well. The steady-state heads  $h_1$  and  $h_2$  in observation wells at radii  $r_1$  and  $r_2$  are monitored till the flow rate  $q$  becomes steady.

The rate of radial flow through any cylindrical surface around the pumping well is equal to the amount of water pumped out. Consider such a surface having radius  $r$ , thickness  $dr$  and height  $h$ . The hydraulic gradient is

Area of flow, From Darcy's Law, Arranging and integrating,  
Field Tests for Permeability

**Confined Flow Pumping Test:** Artesian conditions can exist in a aquifer of thickness  $D$  confined both above and below by impermeable strata. In this, the drawdown water table is above the upper surface of the aquifer.

For a cylindrical surface at radius  $r$ , Integrating,

**Permeability of Stratified Deposits:** When a soil deposit consists of a number of horizontal layers having different permeabilities, the average value of permeability can be obtained separately for both vertical flow and horizontal flow, as  $k_v$  and  $k_h$  respectively.

Consider a stratified soil having horizontal layers of thickness  $H_1, H_2, H_3$ , etc. with coefficients of permeability  $k_1, k_2, k_3$ , etc.

For vertical flow, The flow rate  $q$  through each layer per unit area is the same. Let  $i$  be the equivalent hydraulic gradient over the total thickness  $H$  and let the hydraulic gradients in the layers be  $i_1, i_2, i_3$ , etc. respectively.

where  $k_v$  = Average vertical permeability

The total head drop  $h$  across the layers is

**Horizontal flow:** When the flow is horizontal, the hydraulic gradient is the same in each layer, but the quantity of flow is different in each layer. The total flow is

Considering unit width normal to the cross-section plane,

Worked Examples missing  
of permeability of the sand.

A rectangular soil element is shown with dimensions  $dx$  and  $dz$  in the plane, and thickness  $dy$  perpendicular to this plane. Consider planar flow into the rectangular soil element.

In the  $x$ -direction, the net amount of the water entering and leaving the element is

Similarly in the  $z$ -direction, the difference between the water inflow and outflow is

For a two-dimensional steady flow of pore water, any imbalance in flows into and out of an element in the  $z$ -direction must be compensated by a corresponding opposite imbalance in the  $x$ -direction. Combining the above, and dividing by  $dx \cdot dy \cdot dz$ , the continuity equation is expressed as

From Darcy's law,  $q = -k \frac{dh}{dr}$ , where  $h$  is the head causing flow.

When the continuity equation is combined with Darcy's law, the equation for flow is expressed as:

For an isotropic material in which the permeability is the same in all directions (i.e.  $k_x = k_z$ ), the flow equation is

This is the Laplace equation governing two-dimensional steady state flow. It can be solved graphically, analytically, numerically, or analogically. For the more general situation involving three-dimensional steady flow, Laplace equation becomes:

One-dimensional Flow

For this, the Laplace Equation is

Integrating twice, a general solution is obtained.

The values of constants can be determined from the specific boundary conditions.

As shown, at  $x = 0$ ,  $h = H$ , and at  $x = L$ ,  $h = 0$ . Substituting and solving,  $c_2 = H$ ,

The specific solution for flow in the above permeameter is

which states that head is dissipated in a linearly uniform manner over the entire length of the permeameter.

Two-dimensional Flow

Flow Nets : Graphical form of solutions to Laplace equation for two-dimensional seepage can be presented as flow nets. Two orthogonal sets of curves form a flow net:

Equipotential lines connecting points of equal total head  $h$

Flow lines indicating the direction of seepage down a hydraulic gradient

Two flow lines can never meet and similarly, two equipotential lines can never meet. The space between two adjacent flow lines is known as a flow channel, and the figure formed on the flownet between any two adjacent flow lines and two adjacent equipotential lines is referred to as a field. Seepage through an embankment dam is shown.

Calculation of flow in a channel : If standpipe piezometers were inserted into the ground with their tips on a single equipotential line, then the water would rise to the same level in each standpipe. The pore pressures would be different because of their different elevations. There can be no flow along an equipotential line as there is no hydraulic gradient.

Consider a field of length  $L$  within a flow channel. There is a fall of total head  $= h$ . The average hydraulic gradient is

As the flow lines are  $b$  apart and considering unit length perpendicular to field, the flow rate is

There is an advantage in sketching flow nets in the form of curvilinear 'squares' so that a circle can be inscribed within each four-sided figure bounded by two equipotential lines and two flow lines.

In such a square,  $b = L$ , and the flow rate is obtained as  $q = k \cdot h$

Thus the flow rate through such a flow channel is the permeability  $k$  multiplied by the uniform interval  $= h$  between adjacent equipotential lines.

Calculation of total flow

For a complete problem, the flow net can be drawn with the overall head drop  $h$  divided into  $N_d$  so that  $\Delta h = h / N_d$ .

If  $N_f$  is the no. of flow channels, then the total flow rate is

Procedure for Drawing Flow Nets: At every point  $(x, z)$  where there is flow, there will be a value of head  $h(x, z)$ . In order to represent these values, contours of equal head are drawn.

A flow net is to be drawn by trial and error. For a given set of boundary conditions, the flow net will remain the same even if the direction of flow is reversed. Flow nets are constructed such that the head lost between successive equipotential lines is the same, say  $\Delta h$ . It is useful in visualising the flow in a soil to plot the flow lines, as these are lines that are tangential to the flow at any given point. The steps of construction are:

1. Mark all boundary conditions, and draw the flow cross section to some

convenient scale. 2. Draw a coarse net which is consistent with the boundary conditions and which has orthogonal equipotential and flow lines. As it is usually easier to visualise the pattern of flow, start by drawing the flow lines first.

3. Modify the mesh such that it meets the conditions outlined above and the fields between adjacent flow lines and equipotential lines are 'square'.

4. Refine the flow net by repeating step 3.

The most common boundary conditions are: a) A submerged permeable soil boundary is an equipotential line. This could have been determined by considering imaginary standpipes placed at the soil boundary, as for every point the water level in the standpipe would be the same as the water level. (Such a boundary is marked as CD and EF in the following figure.)

(b) The boundary between permeable and impermeable soil materials is a flow line (This is marked as AB in the same figure). (c) Equipotential lines intersecting a phreatic surface do so at equal vertical intervals.

Uses of Flow Nets: The graphical properties of a flow net can be used in obtaining solutions for many seepage problems such as:

1. Estimation of seepage losses from reservoirs: It is possible to use the flow net in the transformed space to calculate the flow underneath the dam.

2. Determination of uplift pressures below dams: From the flow net, the pressure head at any point at the base of the dam can be determined. The uplift pressure distribution along the base can be drawn and then summed up.

3. Checking the possibility of piping beneath dams: At the toe of a dam when the upward exit hydraulic gradient approaches unity, boiling condition can occur leading to erosion in soil and consequent piping. Many dams on soil foundations have failed because of a sudden formation of a piped shaped discharge channel. As the stored water rushes out, the channel widens and catastrophic failure results. This is also often referred to as piping failure.