CHAPTER 2
SOIL FORMATION AND CHARACTERIZATION

2.1 INTRODUCTION
The word ‘soil’ has different meanings for different professions. To the agriculturist, soil is the top thin layer of earth within which organic forces are predominant and which is responsible for the support of plant life. To the geologist, soil is the material in the top thin zone within which roots occur. From the point of view of an engineer, soil includes all earth materials, organic and inorganic, occurring in the zone overlying the rock crust.

The behavior of a structure depends upon the properties of the soil materials on which the structure rests. The properties of the soil materials depend upon the properties of the rocks from which they are derived. A brief discussion of the parent rocks is, therefore, quite essential in order to understand the properties of soil materials.

2.2 ROCK CLASSIFICATION
Rock can be defined as a compact, semi-hard to hard mass of natural material composed of one or more minerals. The rocks that are encountered at the surface of the earth or beneath, are commonly classified into three groups according to their modes of origin. They are igneous, sedimentary and metamorphic rocks.

Igneous rocks are considered to be the primary rocks formed by the cooling of molten magmas, or by the recrystallization of older rocks under heat and pressure great enough to render them fluid. They have been formed on or at various depths below the earth surface. There are two main classes of igneous rocks. They are:
1. Extrusive (poured out at the surface), and
2. Intrusive (large rock masses which have not been formed in contact with the atmosphere).
Initially both classes of rocks were in a molten state. Their present state results directly from
the way in which they solidified. Due to violent volcanic eruptions in the past, some of the molten
materials were emitted into the atmosphere with gaseous extrusions. These cooled quickly and
eventually fell on the earth’s surface as volcanic ash and dust. Extrusive rocks are distinguished, in
general, by their glass-like structure.

Intrusive rocks, cooling and solidifying at great depths and under pressure containing
entrapped gases, are wholly crystalline in texture. Such rocks occur in masses of great extent, often
going to unknown depths. Some of the important rocks that belong to the igneous group are granite
and basalt. Granite is primarily composed of feldspar, quartz and mica and possesses a massive
structure. Basalt is a dark-colored fine-grained rock. It is characterized by the predominance of
plagioclase, the presence of considerable amounts of pyroxene and some olivine and the absence of
quartz. The color varies from dark-grey to black. Both granite and basalt are used as building
stones.

When the products of the disintegration and decomposition of any rock type are
transported, redeposited, and partly or fully consolidated or cemented into a new rock type, the
resulting material is classified as a sedimentary rock. The sedimentary rocks generally are
formed in quite definitely arranged beds, or strata, which can be seen to have been horizontal at
one time although sometimes displaced through angles up to 90 degrees. Sedimentary rocks are
generally classified on the basis of grain size, texture and structure. From an engineering point of
view, the most important rocks that belong to the group are sandstones, limestones, and shales.

Rocks formed by the complete or incomplete recrystallization of igneous or sedimentary
rocks by high temperatures, high pressures, and/or high shearing stresses are metamorphic rocks.
The rocks so produced may display features varying from complete and distinct foliation of a
crystalline structure to a fine fragmentary partially crystalline state caused by direct compressive
stress, including also the cementation of sediment particles by siliceous matter. Metamorphic rocks
formed without intense shear action have a massive structure. Some of the important rocks that
belong to this group are gneiss, schist, slate and marble. The characteristic feature of gneiss is its
structure, the mineral grains are elongated, or platy, and banding prevails. Generally gneiss is a
good engineering material. Schist is a finely foliated rock containing a high percentage of mica.
Depending upon the amount of pressure applied by the metamorphic forces, schist may be a very
good building material. Slate is a dark colored, platy rock with extremely fine texture and easy
cleavage. Because of this easy cleavage, slate is split into very thin sheets and used as a roofing
material. Marble is the end product of the metamorphism of limestone and other sedimentary rocks
composed of calcium or magnesium carbonate. It is very dense and exhibits a wide variety of
colors. In construction, marble is used for facing concrete or masonry exterior and interior walls
and floors.

Rock Minerals

It is essential to examine the properties of the rock forming minerals since all soils are derived
through the disintegration or decomposition of some parent rock. A ‘mineral’ is a natural inorganic
substance of a definite structure and chemical composition. Some of the very important physical
properties of minerals are crystal form, color, hardness, cleavage, luster, fracture, and specific
gravity. Out of these only two, specific gravity and hardness, are of foundation engineering interest.
The specific gravity of the minerals affects the specific gravity of soils derived from them. The
specific gravity of most rock and soil forming minerals varies from 2.50 (some feldspars) and 2.65
(quartz) to 3.5 (augite or olivine). Gypsum has a smaller value of 2.3 and salt (NaCl) has 2.1. Some
iron minerals may have higher values, for instance, magnetite has 5.2.

It is reported that about 95 percent of the known part of the lithosphere consists of igneous
rocks and only 5 percent of sedimentary rocks. Soil formation is mostly due to the disintegration of
igneous rock which may be termed as a parent rock.
Table 2.1  Mineral composition of igneous rocks

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Percent</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>12-20</td>
</tr>
<tr>
<td>Feldspar</td>
<td>50-60</td>
</tr>
<tr>
<td>Ca, Fe and Mg, Silicates</td>
<td>14-17</td>
</tr>
<tr>
<td>Micas</td>
<td>4-8</td>
</tr>
<tr>
<td>Others</td>
<td>7-8</td>
</tr>
</tbody>
</table>

The average mineral composition of igneous rocks is given in Table 2.1. Feldspars are the most common rock minerals, which account for the abundance of clays derived from the feldspars on the earth’s surface. Quartz comes next in order of frequency. Most sands are composed of quartz.

2.3 FORMATION OF SOILS

Soil is defined as a natural aggregate of mineral grains, with or without organic constituents, that can be separated by gentle mechanical means such as agitation in water. By contrast rock is considered to be a natural aggregate of mineral grains connected by strong and permanent cohesive forces. The process of weathering of the rock decreases the cohesive forces binding the mineral grains and leads to the disintegration of bigger masses to smaller and smaller particles. Soils are formed by the process of weathering of the parent rock. The weathering of the rocks might be by mechanical disintegration, and/or chemical decomposition.

**Mechanical Weathering**

Mechanical weathering of rocks to smaller particles is due to the action of such agents as the expansive forces of freezing water in fissures, due to sudden changes of temperature or due to the abrasion of rock by moving water or glaciers. Temperature changes of sufficient amplitude and frequency bring about changes in the volume of the rocks in the superficial layers of the earth’s crust in terms of expansion and contraction. Such a volume change sets up tensile and shear stresses in the rock ultimately leading to the fracture of even large rocks. This type of rock weathering takes place in a very significant manner in arid climates where free, extreme atmospheric radiation brings about considerable variation in temperature at sunrise and sunset.

Erosion by wind and rain is a very important factor and a continuing event. Cracking forces by growing plants and roots in voids and crevasses of rock can force fragments apart.

**Chemical Weathering**

Chemical weathering (decomposition) can transform hard rock minerals into soft, easily erodable matter. The principal types of decomposition are hydration, oxidation, carbonation, desilication and leaching. Oxygen and carbon dioxide which are always present in the air readily combine with the elements of rock in the presence of water.

2.4 GENERAL TYPES OF SOILS

It has been discussed earlier that soil is formed by the process of physical and chemical weathering. The individual size of the constituent parts of even the weathered rock might range from the smallest state (colloidal) to the largest possible (boulders). This implies that all the weathered constituents of a parent rock cannot be termed soil. According to their grain size, soil...
particles are classified as cobbles, gravel, sand, silt and clay. Grains having diameters in the range of 4.75 to 76.2 mm are called gravel. If the grains are visible to the naked eye, but are less than about 4.75 mm in size the soil is described as sand. The lower limit of visibility of grains for the naked eyes is about 0.075 mm. Soil grains ranging from 0.075 to 0.002 mm are termed as silt and those that are finer than 0.002 mm as clay. This classification is purely based on size which does not indicate the properties of fine grained materials.

Residual and Transported Soils
On the basis of origin of their constituents, soils can be divided into two large groups:

1. Residual soils, and
2. Transported soils.

Residual soils are those that remain at the place of their formation as a result of the weathering of parent rocks. The depth of residual soils depends primarily on climatic conditions and the time of exposure. In some areas, this depth might be considerable. In temperate zones residual soils are commonly stiff and stable. An important characteristic of residual soil is that the sizes of grains are indefinite. For example, when a residual sample is sieved, the amount passing any given sieve size depends greatly on the time and energy expended in shaking, because of the partially disintegrated condition.

Transported soils are soils that are found at locations far removed from their place of formation. The transporting agencies of such soils are glaciers, wind and water. The soils are named according to the mode of transportation. Alluvial soils are those that have been transported by running water. The soils that have been deposited in quiet lakes, are lacustrine soils. Marine soils are those deposited in sea water. The soils transported and deposited by wind are aeolian soils. Those deposited primarily through the action of gravitational force, as in land slides, are colluvial soils. Glacial soils are those deposited by glaciers. Many of these transported soils are loose and soft to a depth of several hundred feet. Therefore, difficulties with foundations and other types of construction are generally associated with transported soils.

Organic and Inorganic Soils
Soils in general are further classified as organic or inorganic. Soils of organic origin are chiefly formed either by growth and subsequent decay of plants such as peat, or by the accumulation of fragments of the inorganic skeletons or shells of organisms. Hence a soil of organic origin can be either organic or inorganic. The term organic soil ordinarily refers to a transported soil consisting of the products of rock weathering with a more or less conspicuous admixture of decayed vegetable matter.

Names of Some Soils that are Generally Used in Practice
Bentonite is a clay formed by the decomposition of volcanic ash with a high content of montmorillonite. It exhibits the properties of clay to an extreme degree.

Varved Clays consist of thin alternating layers of silt and fat clays of glacial origin. They possess the undesirable properties of both silt and clay. The constituents of varved clays were transported into fresh water lakes by the melted ice at the close of the ice age.

Kaolin, China Clay are very pure forms of white clay used in the ceramic industry.

Boulder Clay is a mixture of an unstratified sedimented deposit of glacial clay, containing unsorted rock fragments of all sizes ranging from boulders, cobbles, and gravel to finely pulverized clay material.
**Calcareous Soil** is a soil containing calcium carbonate. Such soil effervesces when tested with weak hydrochloric acid.

**Marl** consists of a mixture of calcareous sands, clays, or loam.

**Hardpan** is a relatively hard, densely cemented soil layer, like rock which does not soften when wet. Boulder clays or glacial till is also sometimes named as hardpan.

**Caliche** is an admixture of clay, sand, and gravel cemented by calcium carbonate deposited from ground water.

**Peat** is a fibrous aggregate of finer fragments of decayed vegetable matter. Peat is very compressible and one should be cautious when using it for supporting foundations of structures.

**Loam** is a mixture of sand, silt, and clay.

**Loess** is a fine-grained, air-borne deposit characterized by a very uniform grain size, and high void ratio. The size of particles ranges between about 0.01 to 0.05 mm. The soil can stand deep vertical cuts because of slight cementation between particles. It is formed in dry continental regions and its color is yellowish light brown.

**Shale** is a material in the state of transition from clay to slate. Shale itself is sometimes considered a rock but, when it is exposed to the air or has a chance to take in water it may rapidly decompose.

### 2.5 SOIL PARTICLE SIZE AND SHAPE

The size of particles as explained earlier, may range from gravel to the finest size possible. Their characteristics vary with the size. Soil particles coarser than 0.075 mm are visible to the naked eye or may be examined by means of a hand lens. They constitute the coarser fractions of the soils. Grains finer than 0.075 mm constitute the finer fractions of soils. It is possible to distinguish the grains lying between 0.075 mm and 2 \( \mu \) (1 \( \mu \) = 1 micron = 0.001 mm) under a microscope. Grains having a size between 2 \( \mu \) and 0.1 \( \mu \) can be observed under a microscope but their shapes cannot be made out. The shape of grains smaller than 1 \( \mu \) can be determined by means of an electron microscope. The molecular structure of particles can be investigated by means of X-ray analysis.

The coarser fractions of soils consist of gravel and sand. The individual particles of gravel, which are nothing but fragments of rock, are composed of one or more minerals, whereas sand grains contain mostly one mineral which is quartz. The individual grains of gravel and sand may be angular, subangular, sub-rounded, rounded or well-rounded as shown in Fig. 2.1. Gravel may contain grains which may be flat. Some sands contain a fairly high percentage of mica flakes that give them the property of elasticity.

Silt and clay constitute the finer fractions of the soil. Any one grain of this fraction generally consists of only one mineral. The particles may be angular, flake-shaped or sometimes needle-like.

Table 2.2 gives the particle size classification systems as adopted by some of the organizations in the USA. The Unified Soil Classification System is now almost universally accepted and has been adopted by the American Society for Testing and Materials (ASTM).

### Specific Surface

Soil is essentially a particulate system, that is, a system in which the particles are in a fine state of subdivision or dispersion. In soils, the dispersed or the solid phase predominates and the dispersion medium, soil water, only helps to fill the pores between the solid particles. The significance of the concept of dispersion becomes more apparent when the relationship of surface to particle size is considered. In the case of silt, sand and larger size particles the ratio of the area of surface of the particles to the volume of the sample is relatively small. This ratio becomes increasingly large as
size decreases from 2 μ which is the upper limit for clay-sized particles. A useful index of relative importance of surface effects is the specific surface of grain. The specific surface is defined as the total area of the surface of the grains expressed in square centimeters per gram or per cubic centimeter of the dispersed phase.

The shape of the clay particles is an important property from a physical point of view. The amount of surface per unit mass or volume varies with the shape of the particles. Moreover, the amount of contact area per unit surface changes with shape. It is a fact that a sphere has the smallest surface area per unit volume whereas a plate exhibits the maximum. Ostwald (1919) has emphasized the importance of shape in determining the specific surface of colloidal systems. Since disc-shaped particles can be brought more in intimate contact with each other, this shape has a pronounced effect upon the mechanical properties of the system. The interparticle forces between the surfaces of particles have a significant effect on the properties of the soil mass if the particles in the media belong to the clay fraction. The surface activity depends not only on the specific surface but also on the chemical and mineralogical composition of the solid particles. Since clay particles

Table 2.2  Particle size classification by various systems

<table>
<thead>
<tr>
<th>Name of the organization</th>
<th>Particle size (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Gravel</td>
</tr>
<tr>
<td>Massachusetts Institute of Technology (MIT)</td>
<td>&gt; 2</td>
</tr>
<tr>
<td>US Department of Agriculture (USDA)</td>
<td>&gt; 2</td>
</tr>
<tr>
<td>American Association of State Highway and Transportation Officials (AASHTO)</td>
<td>76.2 to 2</td>
</tr>
<tr>
<td>Unified Soil Classification System, US Bureau of Reclamation, US Army Corps of Engineers and American Society for Testing and Materials</td>
<td>76.2 to 4.75</td>
</tr>
</tbody>
</table>
are the active portions of a soil because of their high specific surface and their chemical constitution, a discussion on the chemical composition and structure of minerals is essential.

### 2.6 COMPOSITION OF CLAY MINERALS

The word ‘clay’ is generally understood to refer to a material composed of a mass of small mineral particles which, in association with certain quantities of water, exhibits the property of plasticity. According to the clay mineral concept, clay materials are essentially composed of extremely small crystalline particles of one or more members of a small group of minerals that are commonly known as clay minerals. These minerals are essentially hydrous aluminum silicates, with magnesium or iron replacing wholly or in part for the aluminum, in some minerals. Many clay materials may contain organic material and water-soluble salts. Organic materials occur either as discrete particles of wood, leaf matter, spores, etc., or they may be present as organic molecules adsorbed on the surface of the clay mineral particles. The water-soluble salts that are present in clay materials must have been entrapped in the clay at the time of accumulation or may have developed subsequently as a consequence of ground water movement and weathering or alteration processes.

Clays can be divided into three general groups on the basis of their crystalline arrangement and it is observed that roughly similar engineering properties are connected with all the clay minerals belonging to the same group. An initial study of the crystal structure of clay minerals leads to a better understanding of the behavior of clays under different conditions of loading. Table 2.3 gives the groups of minerals and some of the important minerals under each group.

### 2.7 STRUCTURE OF CLAY MINERALS

Clay minerals are essentially crystalline in nature though some clay minerals do contain material which is non-crystalline (for example allophane). Two fundamental building blocks are involved in the formation of clay mineral structures. They are:

1. Tetrahedral unit.
2. Octahedral unit.

The tetrahedral unit consists of four oxygen atoms (or hydroxyls, if needed to balance the structure) placed at the apices of a tetrahedron enclosing a silicon atom which combines together to form a shell-like structure with all the tips pointing in the same direction. The oxygen at the bases of all the units lie in a common plane.

Each of the oxygen ions at the base is common to two units. The arrangement is shown in Fig. 2.2. The oxygen atoms are negatively charged with two negative charges each and the silicon with four positive charges. Each of the three oxygen ions at the base shares its charges with the

<table>
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<tr>
<th>Table 2.3 Clay minerals</th>
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<tbody>
<tr>
<td><strong>Name of mineral</strong></td>
</tr>
<tr>
<td>I. Kaolin group</td>
</tr>
<tr>
<td>1. Kaolinite</td>
</tr>
<tr>
<td>2. Halloysite</td>
</tr>
<tr>
<td>II. Montmorillonite group</td>
</tr>
<tr>
<td>Montmorillonite</td>
</tr>
<tr>
<td>III. Illite group</td>
</tr>
<tr>
<td>Illite</td>
</tr>
</tbody>
</table>
adjacent tetrahedral unit. The sharing of charges leaves three negative charges at the base per tetrahedral unit and this along with two negative charges at the apex makes a total of 5 negative charges to balance the 4 positive charges of the silicon ion. The process of sharing the oxygen ions at the base with neighboring units leaves a net charge of \(-1\) per unit.

The second building block is an octahedral unit with six hydroxyl ions at apices of an octahedral enclosing an aluminum ion at the center. Iron or magnesium ions may replace aluminum ions in some units. These octahedral units are bound together in a sheet structure with each hydroxyl ion common to three octahedral units. This sheet is sometimes called as gibbsite sheet. The Al ion has 3 positive charges and each hydroxyl ion divides its \(-1\) charge with two other neighboring units. This sharing of negative charge with other units leaves a total of 2 negative charges per unit \((1/3) \times 6\). The net charge of a unit with an aluminum ion at the center is \(+1\). Fig. 2.3 gives the structural arrangements of the units. Sometimes, magnesium replaces the aluminum atoms in the octahedral units in this case, the octahedral sheet is called a brucite sheet.

**Formation of Minerals**

The combination of two sheets of silica and gibbsite in different arrangements and conditions lead to the formation of different clay minerals as given in Table 2.3. In the actual formation of the sheet silicate minerals, the phenomenon of *isomorphous substitution* frequently occurs. Isomorphous (meaning same form) substitution consists of the substitution of one kind of atom for another.

**Kaolinite Mineral**

This is the most common mineral of the kaolin group. The building blocks of gibbsite and silica sheets are arranged as shown in Fig. 2.4 to give the structure of the kaolinite layer. The structure is composed of a single tetrahedral sheet and a single alumina octahedral sheet combined in units so that the tips of the silica tetrahedrons and one of the layers of the octahedral sheet form a common layer. All the tips of the silica tetrahedrons point in the same direction and towards the center of the unit made of the silica and octahedral sheets. This gives rise to strong ionic bonds between the silica and gibbsite sheets. The thickness of the layer is about 7 Å (one angstrom = \(10^{-8}\) cm) thick. The kaolinite mineral is formed by stacking the layers one above the other with the base of the silica sheet bonding to hydroxyls of the gibbsite sheet by hydrogen bonding. Since hydrogen bonds are comparatively strong, the kaolinite

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**Figure 2.2** Basic structural units in the silicon sheet (Grim, 1959)
crystals consist of many sheet stackings that are difficult to dislodge. The mineral is therefore, stable, and water cannot enter between the sheets to expand the unit cells. The lateral dimensions of kaolinite particles range from 1000 to 20,000 Å and the thickness varies from 100 to 1000 Å. In the kaolinite mineral there is a very small amount of isomorphous substitution.

**Halloysite Mineral**

Halloysite minerals are made up of successive layers with the same structural composition as those composing kaolinite. In this case, however, the successive units are randomly packed and may be separated by a single molecular layer of water. The dehydration of the interlayers by the removal of the water molecules leads to changes in the properties of the mineral. An important structural feature of halloysite is that the particles appear to take tubular forms as opposed to the platy shape of kaolinite.

![Figure 2.3 Basic structural units in octahedral sheet (Grim, 1959)](image)

![Figure 2.4 Structure of kaolinite layer](image)
Montmorillonite Mineral

Montmorillonite is the most common mineral of the montmorillonite group. The structural arrangement of this mineral is composed of two silica tetrahedral sheets with a central alumina octahedral sheet. All the tips of the tetrahedra point in the same direction and toward the center of the unit. The silica and gibbsite sheets are combined in such a way that the tips of the tetrahedrons of each silica sheet and one of the hydroxyl layers of the octahedral sheet form a common layer. The atoms common to both the silica and gibbsite layer become oxygen instead of hydroxyls. The thickness of the silica-gibbsite-silica unit is about 10 Å (Fig. 2.5). In stacking these combined units one above the other, oxygen layers of each unit are adjacent to oxygen of the neighboring units with a consequence that there is a very weak bond and an excellent cleavage between them. Water can enter between the sheets, causing them to expand significantly and thus the structure can break into 10 Å thick structural units. Soils containing a considerable amount of montmorillonite minerals will exhibit high swelling and shrinkage characteristics. The lateral dimensions of montmorillonite particles range from 1000 to 5000 Å with thickness varying from 10 to 50 Å. Bentonite clay belongs to the montmorillonite group. In montmorillonite, there is isomorphous substitution of magnesium and iron for aluminum.

Illite

The basic structural unit of illite is similar to that of montmorillonite except that some of the silicons are always replaced by aluminum atoms and the resultant charge deficiency is balanced by potassium ions. The potassium ions occur between unit layers. The bonds with the nonexchangeable K\(^+\) ions are weaker than the hydrogen bonds, but stronger than the water bond of montmorillonite. Illite, therefore, does not swell as much in the presence of water as does montmorillonite. The lateral dimensions of illite clay particles are about the same as those of montmorillonite, 1000 to 5000 Å, but the thickness of illite particles is greater than that of montmorillonite particles, 50 to 500 Å. The arrangement of silica and gibbsite sheets are as shown in Fig. 2.6.

2.8 CLAY PARTICLE-WATER RELATIONS

The behavior of a soil mass depends upon the behavior of the discrete particles composing the mass and the pattern of particle arrangement. In all these cases water plays an important part. The
behavior of the soil mass is profoundly influenced by the inter-particle-water relationships, the ability of the soil particles to adsorb exchangeable cations and the amount of water present.

**Adsorbed Water**

The clay particles carry a net negative charge on their surface. This is the result of both isomorphous substitution and of a break in the continuity of the structure at its edges. The intensity of the charge depends to a considerable extent on the mineralogical character of the particle. The physical and chemical manifestations of the surface charge constitute the surface activity of the mineral. Minerals are said to have high or low surface activity, depending on the intensity of the surface charge. As pointed out earlier, the surface activity depends not only on the specific surface but also on the chemical and mineralogical composition of the solid particle. The surface activity of sand, therefore, will not acquire all the properties of a true clay, even if it is ground to a fine powder. The presence of water does not alter its properties of coarser fractions considerably excepting changing its unit weight. However, the behavior of a saturated soil mass consisting of fine sand might change under dynamic loadings. This aspect of the problem is not considered here. This article deals only with clay particle-water relations.

In nature every soil particle is surrounded by water. Since the centers of positive and negative charges of water molecules do not coincide, the molecules behave like dipoles. The negative charge on the surface of the soil particle, therefore, attracts the positive (hydrogen) end of the water molecules. The water molecules are arranged in a definite pattern in the immediate vicinity of the boundary between solid and water. More than one layer of water molecules sticks on the surface with considerable force and this attractive force decreases with the increase in the distance of the water molecule from the surface. The electrically attracted water that surrounds the clay particle is known as the *diffused double-layer of water*. The water located within the zone of influence is known as the *adsorbed layer* as shown in Fig. 2.7. Within the zone of influence the physical properties of the water are very different from those of free or normal water at the same temperature. Near the surface of the particle the water has the property of a solid. At the middle of the layer it resembles a very viscous liquid and beyond the zone of influence, the properties of the water become normal. The adsorbed water affects the behavior of clay particles when subjected to external stresses, since it comes between the particle surfaces. To drive off the adsorbed water, the clay particle must be heated to more than 200 °C, which would indicate that the bond between the water molecules and the surface is considerably greater than that between normal water molecules.

![Figure 2.6 Structure of illite layer](image-url)
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The adsorbed film of water on coarse particles is thin in comparison with the diameter of the particles. In fine grained soils, however, this layer of adsorbed water is relatively much thicker and might even exceed the size of the grain. The forces associated with the adsorbed layers therefore play an important part in determining the physical properties of the very fine-grained soils, but have little effect on the coarser soils.

Soils in which the adsorbed film is thick compared to the grain size have properties quite different from other soils having the same grain sizes but smaller adsorbed films. The most pronounced characteristic of the former is their ability to deform plastically without cracking when mixed with varying amounts of water. This is due to the grains moving across one another supported by the viscous interlayers of the films. Such soils are called cohesive soils, for they do not disintegrate with pressure but can be rolled into threads with ease. Here the cohesion is not due to direct molecular interaction between soil particles at the points of contact but to the shearing strength of the adsorbed layers that separate the grains at these points.

**Base Exchange**

Electrolytes dissociate when dissolved in water into positively charged cations and negatively charged anions. Acids break up into cations of hydrogen and anions such as Cl or SO$_4$. Salts and bases split into metallic cations such as Na, K or Mg, and nonmetallic anions. Even water itself is an electrolyte, because a very small fraction of its molecules always dissociates into hydrogen ions H$^+$ and hydroxyl ions OH$^-$. These positively charged H$^+$ ions migrate to the surface of the negatively charged particles and form what is known as the adsorbed layer. These H$^+$ ions can be replaced by other cations such as Na, K or Mg. These cations enter the adsorbed layers and constitute what is termed as an adsorption complex. The process of replacing cations of one kind by those of another in an adsorption complex is known as base exchange. By base exchange is meant the capacity of
Table 2.4 Exchange capacity of some clay minerals

<table>
<thead>
<tr>
<th>Mineral group</th>
<th>Exchange capacity (meq per 100 g)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kaolinites</td>
<td>3.8</td>
</tr>
<tr>
<td>Illites</td>
<td>40</td>
</tr>
<tr>
<td>Montmorillonites</td>
<td>80</td>
</tr>
</tbody>
</table>

Table 2.5 Cations arranged in the order of decreasing shear strength of clay

$\text{NH}_4^+ > \text{H}^+ > \text{K}^+ > \text{Fe}^{+++} > \text{Al}^{+++} > \text{Mg}^+ > \text{Ba}^{++} > \text{Ca}^{++} > \text{Na}^+ > \text{Li}^+$

colloidal particles to change the cations adsorbed on their surface. Thus a hydrogen clay (colloid with adsorbed H cations) can be changed to sodium clay (colloid with adsorbed Na cations) by a constant percolation of water containing dissolved Na salts. Such changes can be used to decrease the permeability of a soil. Not all adsorbed cations are exchangeable. The quantity of exchangeable cations in a soil is termed exchange capacity.

The relative exchange capacity of some of the clay minerals is given in Table 2.4.

If one element, such as H, Ca, or Na prevails over the other in the adsorption complex of a clay, the clay is sometimes given the name of this element, for example H-clay or Ca-clay. The thickness and the physical properties of the adsorbed film surrounding a given particle depend to a large extent on the character of the adsorption complex. The films are relatively thick in the case of strongly water-adsorbent cations such as Li$^+$ and Na$^+$ cations but very thin for H$^+$. The films of other cations have intermediate values. Soils with adsorbed Li$^+$ and Na$^+$ cations are relatively more plastic at low water contents and possess smaller shear strength because the particles are separated by a thicker viscous film. The cations in Table 2.5 are arranged in the order of decreasing shear strength of clay.

Sodium clays in nature are a product either of the deposition of clays in sea water or of their saturation by saltwater flooding or capillary action. Calcium clays are formed essentially by fresh water sediments. Hydrogen clays are a result of prolonged leaching of a clay by pure or acidic water, with the resulting removal of all other exchangeable bases.

2.9 SOIL MASS STRUCTURE

The orientation of particles in a mass depends on the size and shape of the grains as well as upon the minerals of which the grains are formed. The structure of soils that is formed by natural deposition can be altered by external forces. Figure 2.8 gives the various types of structures of soil. Fig. 2.8(a) is a single grained structure which is formed by the settlement of coarse grained soils in suspension in water. Fig. 2.8(b) is a flocculent structure formed by the deposition of the fine soil fraction in water. Fig. 2.8(c) is a honeycomb structure which is formed by the disintegration of a flocculent structure under a superimposed load. The particles oriented in a flocculent structure will have edge-to-face contact as shown in Fig 2.8(d) whereas in a
honeycomb structure, the particles will have face-to-face contact as shown in Fig. 2.8(e). Natural clay sediments will have more or less flocculated particle orientations. Marine clays generally have a more open structure than fresh water clays. Figs. 2.8(f) and (g) show the schematic views of salt water and fresh water deposits.

Figure 2.8 Schematic diagrams of various types of structures (Lambe, 1958a)