Chapter

2

Soil Physical Environment

The physical properties of soil play many significant roles in determining the physical environment of soil and its suitability for crop production. Soil is a three-phase system: solid (mineral particles and organic matter), liquid (water), and gas (air). The physical properties of soil depend on the amount, shape, size, composition, and arrangement of inorganic or mineral soil particles and the amount and nature of organic soil particles (organic matter) and their quantitative relation with the other two components of soil, viz water, and air. The behaviour of soil, like mechanical supporting power, moisture retention and supply capacity to plants, drainage, aeration, the penetrability of roots, retention of plant nutrients, and availability, are intimately related to the physical properties of soil.

2.1 SOIL TEXTURE

The inorganic or mineral soil component comprises two types of soil particles: primary and secondary particles. **Primary particles or soil separates** are discrete units that cannot be further subdivided, while **secondary particles** consist of primary particles and can be subdivided by chemical or mechanical dispersion. The diameter of individual soil particles varies from boulders (1 m) to sub-microscopic clay (10^{-6} m). Gravel, cobbles, boulders, and other coarse fragments larger than 2 mm in diameter may affect the behaviour of soil but are not considered as a part of the fine earth fractions to which the term soil texture applies.

Based on equivalent diameter, primary particles or soil separates smaller than 2 mm diameter are usually divided into three groups: (i) sand, (ii) silt, and (iii) clay. **Soil texture** is defined as the relative proportions of sand, silt, and clay fractions in a soil. The soil texture is almost a permanent intrinsic property of soil that influences several soil behaviours such as aeration status, water retention capacity, drainage facility, ease of tillage, etc.

2.1.1 Particle Size Classification

There are numerous systems of classification of soil separates into different size limits. The two classification systems most widely used by soil scientists and agronomists are the U.S. Department of Agriculture (USDA) and the International Society of Soil Science (ISSS), renamed the International Union of Soil Science (IUSS) as shown in Fig. 2.1.

Sand: It is the coarsest fraction and constitutes the skeleton of a soil body. As sand particles are relatively large, the pore spaces between them are relatively large, thus facilitating free drainage of water and entry of air into the soil. Sand particles having relatively less specific surface area (the surface area per unit volume or mass) can hold little water.

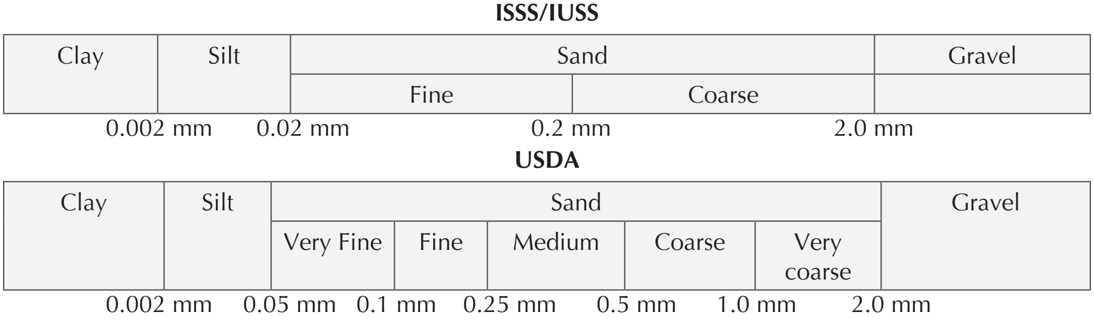


Fig. 2.1: Classification of soil particles according to their size

Silt: It is an intermediate-size fraction between sand and clay. The properties of coarser silt fraction resemble that of sand, while the properties of finer silt approach that of clay. Silt particles being relatively smaller in size possess more specific surface area than the sand.

Clay: It is the finest and most reactive fraction of the soil separates. While sand and silt particles constitute the skeleton, the clay fraction is analogous to the flesh of the soil body. Very fine clay particles (< 0.001 mm or 1 µm) behave like a colloid. Because of their higher specific surface area and high charge density, they can adsorb water, nutrient ions, and pollutants. Clay particles impart swell/shrink, plasticity, and stickiness to the soil.

Three broad groups of soil textural classes are recognized: sandy soils, clayey soils, and loamy soils. Each group is subdivided into **specific textural classes** that give the idea of the size distribution of particles and eventually indicate the nature of soil's physical properties. In sandy soils, sand separates comprise at least 70% of the material by weight (and less than 15% clay). Two specific textural classes in this group are sands and loamy sands. The loamy soils group consists of many subdivisions: sandy loam, loam, silt loam, silt, sandy clay loam, silty clay loam, and clay loam. An ideal loam is a mixture of sand, silt, and clay that exhibits the properties of those separates in about equal proportions. To qualify for the modifier sandy or silt, the soil must contain at least 40–45% of that separate, respectively, while the clay modifier requires at least 20% clay. In the clayey soil group, the clay separates comprise at least 35% of the materials by weight. Three specific textural classes are recognized in this group: sandy clay, silty clay, and clay. Although, clayey soils are more porous, require more power to plough than sandy soils. At low moisture levels, clayey soils are hard and coherent due to the cementation effect of clay particles; on the other hand, at high moisture levels, these soils are very sticky and plastic because of higher surface area.

2.2 SPECIFIC SURFACE AREA

It can be expressed as the total surface area per unit mass (cm^2/g) or per unit volume (cm^2/cm^3). Several soil properties are intimately associated with the specific surface area of soil, which include cation exchange capacity, retention, and movement of various chemicals, swell shrink capacity, plasticity, etc. The surface area depends on the size and shape of the particle. It increases logarithmically with a decrease in particle size (Table 2.1). Again, it differs with the type of clay, for example, montmorillonite (550–650 m^2/g) has a surface area many times greater than kaolinite (5–30 m^2/g) and illite (70–175 m^2/g).

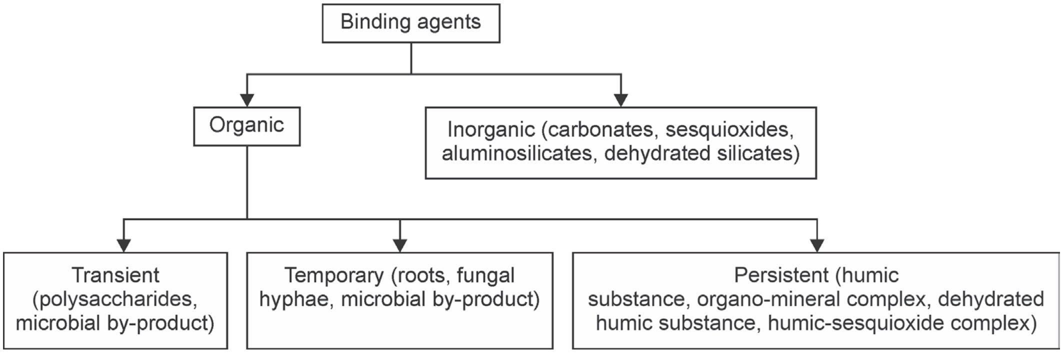
Table 2.1: Specific surface area of soil particles

Size fraction	***Diameter (mm)***	***Specific surface area (cm^2/g)***
Coarse sand	2–0.2	45
Fine sand	0.2–0.02	446
Silt	0.02–0.002	4,458
Ultra-clay	10^{-4}–10^{-6}	9.89×10^6

2.3 SOIL STRUCTURE

The primary particles (sand, silt, and clay) generally cluster together and are stabilized by some cementing agents to form complex and irregular patterns of secondary particles called **aggregates** or **peds**. The term **soil structure** refers to the arrangement of primary and secondary particles into a definite structural pattern. Various types of structural peds occur in soils, even within different horizons of a particular soil profile. Soil structure is characterized based on the shape and arrangement, size, and stability of the peds. Soil structure notably influences several soil-physical properties such as water retention and movement, heat transfer, aeration, and porosity. Various soil management practices such as tillage, application of manure, fertilizer, lime, and irrigation water may lead to changes in soil structure, particularly in the surface soils.

The aggregate stability depends on the nature of the binding agents involved in cementing the particles together. There are different binding agents at each step moving from clay to macroaggregate. These can be grouped as follows:



The genesis of soil structure is largely a function of:

i. Nature of cations and anions present in the soil
ii. Amount of silt and clay particles
iii. Iron and aluminium oxide content
iv. Nature of organic compounds present in the soil
v. Vegetation, soil fauna, and microorganisms

Alkali soils, characterized by poor structural qualities are changed into favourable ones if the sodium is replaced from the exchange complex by calcium through liming. Phosphate plays a significant role in soil aggregation. Stable aggregate formation cannot take place in sands or silts in the absence of colloids. A higher correlation between clay content and aggregation is observed as the amount of organic matter decreased. Again, at a higher amount of organic matter, the effect of clay in aggregate formation becomes insignificant. A very close relationship exists between the amount of free iron and water-stable aggregates in laterite soil. The impact of fungi and actinomycetes on soil aggregation is temporary and disappears with the bacterial decomposition of mycelia. Aggregate stability seems to be a function of microbial synthesis of cementing agents rather than the physical presence of soil microorganisms. Intermediate products of decomposition, mostly polysaccharides and bacterial gums act in binding the soil particles into aggregates. Humic acid which is a long chain flexible polymer, a fairly stable product of bacterial decomposition is more efficient than fulvic acids in soil aggregation. Besides the cementing action of amino-polyuronides, polysaccharides, protein, and lignin–like colloidal substances, fat, resins, and waxes may arrest the slaking of aggregates by making them waterproof and ultimately increase stability. The possible role of vegetation in soil aggregation is multidimensional. The high state of aggregation under grassland vegetation has been attributed to the desiccating action of the extensive root system, which creates an alternate wetting and drying effect. The stability of soil structure through microbial action is the integrated effect of (i) the mechanical binding of particles by microbial cells and filaments of fungi and actinomycetes, (ii) the cementing action of the products of microbial synthesis, and (iii) the stabilization action of the end products of decomposition (humic acid, colloidal proteins, etc.) acting individually or in combination. The following are some indices used to characterize soil structure:

i. Mean weight diameter (MWD)
ii. Geometric mean diameter (GMW)
iii. Stability index
iv. Structural coefficient or stability coefficient (SC)
v. Dispersion coefficient (DC)
vi. Minimum bulk density
vii. High aeration pore space
viii. Hydraulic conductivity at minimum bulk density
ix. Organic carbon

Ultimately, a good soil structure indicates soil possessing higher pore space and easy percolation of rain/irrigation water. Therefore, hydraulic conductivity at minimum bulk density value is considered the best among the indices.

2.4 BULK DENSITY AND POROSITY

Bulk density can be calculated in two soil conditions—dry and wet soil conditions. The dry bulk density (ρ_b) or apparent density of soil is defined as the mass per unit volume of dry soil. Here, the volume includes both solid volume and pore spaces.

$$\rho_b = \frac{M_s}{V_t} = \frac{M_s}{V_s + V_a}$$

where V_t is the total volume of soil which includes the volume occupied by soil solid (V_s) and by air (V_a) and/or water (V_w), i.e. void volume (V_v). In dry soil, V_w equals zero, then $V_v = V_a$ and $V_t = V_s + V_a$.

From the above equation, it is also obvious that any factor that influences the pore volume will affect the bulk density of soil since no one can manipulate the volume of the soil solid. Thus, the bulk density of soil is influenced by soil texture, structure, organic matter content, and management practices.

Fine-textured soils, such as silt loams, clay loams, and clays generally exhibit lower bulk densities than sandy soils. In fine-textured soil, solid particles are organized in porous granules. Total porosity in fine-textured soils is always higher than coarse-textured soils. Pores in the fine-textured soil having bulk densities ranging from 1.10 to 1.40 Mg m^{-3} are present both between and within the granules, while in coarse-textured sandy soils having bulk densities ranging from 1.40 to 1.75 Mg m^{-3} most the pores are large and a smaller number of pores than fine-textured soil. The structure of soil affects the pore space and pore geometry, thereby the bulk density. Tightly packed soils exhibit higher bulk densities than loosely packed soils. Organic matter itself is lighter than the mineral particles. Again, organic matter helps in the aggregation of soil particles to form granules. As in aggregated soils, pores exist both between and within the granules, so the total pore space will be higher. Thus, soils containing higher amounts of organic matter are always more porous and exhibit lower bulk densities than the soil with less organic matter content. Tillage temporarily makes the surface soil loose and decreases the bulk density of the soil, but in the long run, intense tillage increases bulk density due to the depletion of organic matter by oxidation and breaking of soil aggregates. The bulk density of subsoil is usually higher than the surface soil, not only because of less organic matter content but also due to compaction done in subsoil by running heavy machines for different agricultural operations and the weight of the overlying soil layers.

Porosity refers to the relative volume of voids (pores) and is expressed as a fraction or % of the total volume of soil or the volume of solid. It is expressed in the following ways:

Total Porosity (*f*): Total porosity (*f*) is the ratio of the total volume of pore spaces (volume not occupied by soil solids) to the total volume of soil. It can also be expressed in percentages.

$$f = \frac{V_{\mathrm{v}}}{V_{\mathrm{t}}} = \frac{V_{\mathrm{w}} + V_{\mathrm{a}}}{V_{\mathrm{s}} + V_{\mathrm{w}} + V_{\mathrm{a}}}$$

Total porosity or % pore space values usually vary from 30–60%, i.e. in the fraction 0.30 to 0.60. Concerning function, pores are of two types: **Transmission pores** (> 0.05 mm in diameter) and **storage pores** (< 0.05 mm in diameter). While transmission pores are spaces for the movement of air and drainage of excess water, storage pores are spaces for water retention against gravity. Coarse-textured soils are dominated by transmission pores, while fine-textured soils are dominated by storage pores, and the latter is usually more porous than the former. Like bulk density, total porosity varies among the soils, even in the same soil with various management practices.

When the soil is saturated, pores remain filled with water. As the soil starts to dry, the larger pores empty first. The volume of water retained by the soil at 50 cm tension is called **capillary porosity** (usually indicates the volume of storage pores). The **noncapillary porosity** (usually transmission pores) is calculated as the difference between total and capillary porosity. Both porosities are important for plant growth, as capillary porosity is the volume for plant-available water retention, and noncapillary porosity is the volume for drainage and movement of air (root aeration) within the soil. Good aeration can only occur when the proportion of noncapillary pore to the total porosity should be at least 10%.

2.5 SOIL COMPACTION

Soil compaction refers to the increase in the dry bulk density of soil by applied loads or pressure. In practice, soil compaction is a process leading to the compression of a mass of soil by applied force and physical deformation resulting in a decrease in total as well as transmission pores. Since the compression of solids and water is impossible; the compacting force can only cause the rearrangement or deformation of soil solids. Thus, the decrease in volume is due to compression or expulsion of soil air, which may be expelled or compressed. Therefore, the degree or severity of soil compaction is expressed in relation to bulk density, total porosity, aeration porosity, and void ratio.

Soil compaction is intimately related to agriculture due to its adverse effect on root development and crop yields, and to the environment due to its indirect effect on soil erosion, anaerobiosis, and transport of pollutants in the surface and subsurface flow, and the exchange of gases between soil and atmosphere. Concerning plant root growth, the optimum bulk density for most soils is <1.4 Mg m^{-3}. However, it may vary among soils and crops, e.g. <1.0 Mg m^{-3} for Andisols. Thus, from an agricultural perspective, the objective of soil management is to sustain the bulk density within an optimal range that favours root growth, transmission and retention of water, and gaseous exchange.

2.6 SOIL WATER

Water is an important component of every life. The basic physiological and metabolic processes of plants, like photosynthesis and nutrient absorption cannot take place in the absence of water. Water acts as a solvent for several salts carrying plant nutrients, thus becoming the only transport in soil for the movement of nutrient elements and pollutants. For this unique role, it is imperative to learn the energy status, retention, and movement behaviour of soil water.

From the knowledge of the structure and properties of water, it is obvious that three forces determine the energy status of soil water. First, the adhesion of water to soil solids (matrix) due to **matric** force (responsible for adhesion and capillarity) reduces the energy state of water near the particle surface. Second, the attraction of water molecules to ions and salts due to **osmotic** force reduces the energy state of water in soil solution. The third force is **gravity** acting on soil water, which always pulls water downward (Fig. 2.2).

Measuring the absolute energy level of soil water is a difficult task. However, its movement in the soil can be predicted not by absolute energy level but by the difference in the energy level of soil water relative to that of pure water located at some reference elevation at standard pressure and temperature unaffected by the soil. The difference in energy levels between this pure water in the reference state and soil water is termed **total soil water potential** (potential energy per unit quantity mass/weight/volume is potential). In fact, in practice, the potential is not measured by transporting water as per the definition but rather by measuring some other properties, e.g. elevation, hydrostatic pressure, osmotic pressure, etc. Soil water potential is expressed in units of energy per unit mass (ergs/g or J/kg), unit volume (dynes/cm^2, bar, atmosphere, kilopascals, etc.), or unit weight (cm or m).

The total soil water potential consists of several components:

$$\psi_t = \psi_z + \psi_p + \psi_o + \psi_\Omega + \ldots,$$

where ψ_t, ψ_z, ψ_p, ψ_o and ψ_Ω refer to **total, gravitational, pressure, osmotic and overburden potential**, respectively. Under the unsaturated condition, pressure potential encompasses **matric potential** and **pneumatic potential** (ψ_a) [due to change in pressure of ambient air].

So, for unsaturated soil, $\psi_t = \psi_z + \psi_p + \psi_o + \ldots$
$= \psi_z + \psi_m + \psi_a + \psi_o + \ldots$
$= \psi_h + \psi_o + \ldots$

A sum of gravitational and pressure potential is called **hydraulic potential** or **hydraulic head** (ψ_h).

Under saturated conditions, there is no existence of matric potential and here pressure potential is due to the weight of water of submergence called **submergence potential** (ψ_s).

So, for saturated soil, $\psi_t = \psi_z + \psi_s + \psi_o + \ldots$

The gravitational potential is merely due to the position of soil water with respect to a specified elevation of the point under consideration. The gravitational potential can be measured above or below the arbitrarily chosen reference level. It is positive if the specified point of concern lies above the reference level and is negative when it lies below the reference level.

Pressure potential commonly includes: (i) the positive hydrostatic pressure resulting from the weight of water in saturated soils and aquifers, and (ii) the negative pressure because of the attractive forces between water and soil solids (soil matrix). The positive hydrostatic pressure (as hydrostatic pressure is higher than atmospheric pressure) develops if the point under consideration is below the water table and it is just equal and opposite of gravitational potential; and is measured from the free water surface (Fig. 2.2). It is sometimes referred to as **submergence potential** or **piezometric potential** which corresponds to piezometric head. Pressure potential is the vertical distance between the specified point in the soil and the water surface in the piezometer. ψ_p is zero above or at the level of water in the piezometer. Its value is positive and equal to the depth of the water column above the specified point when it is below the water table.

Matric potential arises due to the attraction of water to solid surfaces above the water table and is always negative since the water attracted by the soil matrix has an energy state lesser than that of pure water (the free energy of pure water is considered zero). It is the component of total soil water potential only in unsaturated soil because under saturated conditions the attraction of water to solid surfaces is fully satisfied and is considered as zero.

This negative pressure potential is referred to as **suction** or **tension**. Matric potential arises due to cohesive force (attraction between water molecules), adhesive force (attraction between solids and water), and surface tension at air-water interfaces. It is also called **capillary potential**. The difference in matric potential between two adjoining zones of

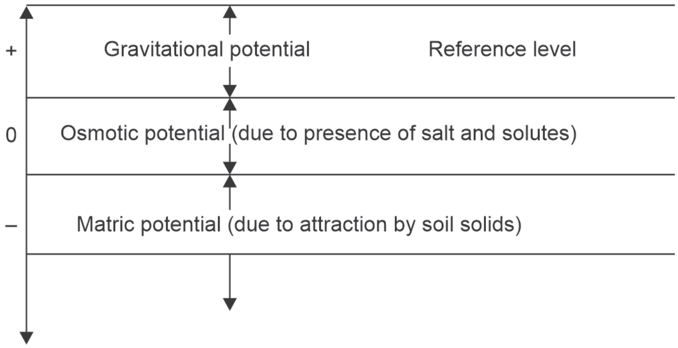


Fig. 2.2: Potential energy of pure water and soil water

soil is the driving force for the movement of water from the moist zone (higher energy state) to the dry zone (lower energy state). This movement may be slow, but is extremely important so far as the water supply to plant roots is concerned.

Osmotic potential arises because of the presence of solutes (inorganic salts or organic compounds) in soil solution. The polar water molecules attract and cluster around each solute ion or molecule and thus reduce the free energy of pure water. The greater the concentration of solute, the lesser will be the osmotic potential. As always, water tends to move to where its energy level lowers. Water moves from a lower concentration (higher osmotic potential) to the zone of higher concentration (lower osmotic potential) by the process of **osmosis** if the two zones are separated by a semipermeable membrane that prevents the movement of solute ions only. During the mass flow, this potential is insignificant as there is no semipermeable membrane in the soil. This potential has a major bearing on the water uptake by plant roots as the root cell membrane acts as a semipermeable membrane between the soil solution and the root cell. In soils of high salt concentration (saline soil), lower osmotic potential (greater negative value) in soil solution than root cells restricts the water absorption by plants. In high saline soils, young seedlings may collapse due to exosmosis (**plasmolysis**), i.e. water movement from root cell to soil solution.

The term tension or suction is used in place of negative potentials in the case of matric and osmotic potential if water is at sub-atmospheric pressure (unsaturated condition) to avoid the negative sign before the values. When negative potentials of unsaturated soils are expressed in terms of equivalent hydraulic head, the values become high enough in the range of -10^5 or -10^6 cm of water. To avoid the use of large values with negative signs, Schofield (1935) introduced a pF scale that is a logarithm to the base 10 of the negative pressure head (tension or suction) in centimetres (h) of the water column.

$$\mathrm{pF} = \log h$$

Thus, a tension or suction head of 1000 cm water on a pF scale is 3.

2.6.1 Soil Moisture Constants

Soil moisture characteristics may be utilized to derive several moisture retention quantities corresponding to the water potential values which are of great practical significance to irrigation and drainage management of soil.

Maximum Retention Capacity: When all the pore spaces are filled with water received from irrigation or rainfall, the soil is said to be saturated with water and at its *maximum retention capacity* or *maximum water holding capacity*. The matric potential at this point of moisture content is nearly the same as that of pure water ($\psi_m \sim 0$). At this point, water is present both in smaller and larger pores which are readily percolated downward under the influence of gravity.

Field Capacity: When a fully saturated soil is allowed to drain freely by gravity, and there is no loss of water by evaporation, after some time, the soil moisture will attain equilibrium, called *field moisture capacity* or *field capacity*. It is the upper limit of soil moisture content when the downward movement of water under the influence of gravity from a pre-saturated soil ceases. It corresponds to the matric soil water potential of –0.1 bar (equivalent to pF 2) for coarse textured and –0.33 bar (equivalent to pF 2.5) for fine textured soil. In the field, the soil moisture at the surface layer attains this steady state generally 2 to 3 days after saturation. At this point, water is present only in capillary pores.

Wilting Coefficient: It corresponds to the moisture content retained by the soil when the matric potential is –15 bar (equivalent to pF 4.19). Below this soil moisture, the plant cannot extract water at a rate sufficient to meet the transpiration demand. Eventually, plant cells lose turgidity and start to exhibit the symptoms of wilting. The *wilting coefficient or permanent wilting point* corresponds to the maximum water content at which plants grown under such conditions get wilted so severely that they cannot regain their turgidity even if exposed to a water-vapour-saturated atmosphere.

Available Water: Plant available water is a part of soil water that soils hold at a potential between field capacity (0.1 or 0.33 bar suction) and wilting coefficient (15 bar suction). Available water held at suction between 0.1 or 0.33 bar and 15 bar is not equally available. Above a threshold suction (1.0 bar for coarser and 5 bars for finer soil), water availability decreases linearly with an increase in soil water suction till it becomes unavailable at 15 bars. Soil texture has a profound influence on field capacity and permanent wilting point, thereby on the available water content.

$$AWC = FC - PWP$$

With the increase in fineness of texture, there is a general increase in available water content from sandy to silt loam soils. Both FC and PWP increase with the increase in the fineness of texture, but the increase in FC is more than the PWP. However, clay soils usually provide less available water than well-aggregated silt loams. Clays tend to have almost the same moisture content at field capacity but have higher moisture content at the permanent wilting point than the silt loam (Fig. 2.3) resulting in less difference.

The available water content of the soil increases with an increase in the organic matter content of mineral soils. This is due to the high water retention capacity of organic matter

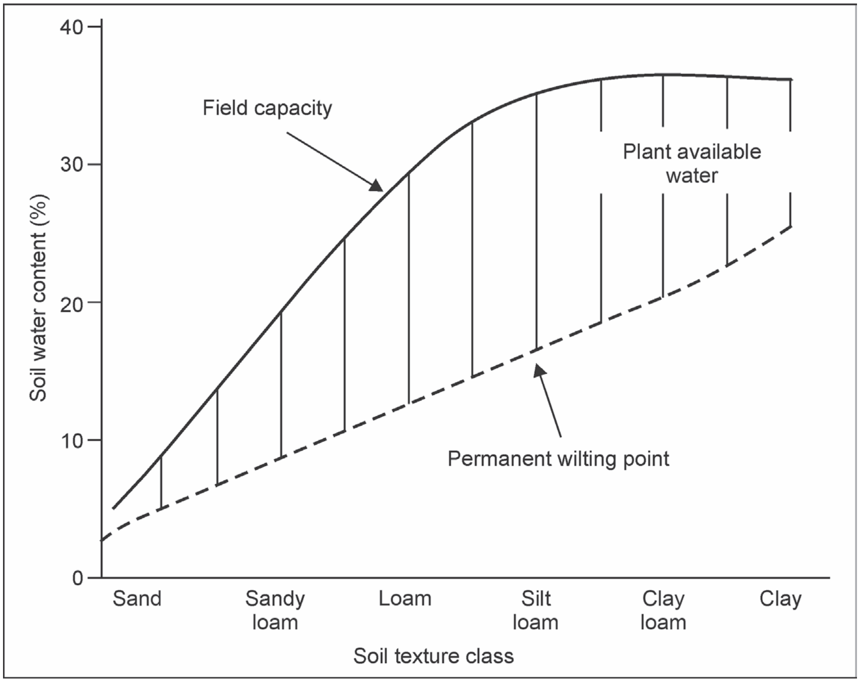


Fig. 2.3: Effect of soil texture on plant available water content

at field capacity rather than an equal volume of mineral matter. The influence of organic matter on soil structure improvement and increase in total porosity indirectly helps to increase the plant available water content of the soil. In mineral soil, an increase in organic matter content increases moisture content at both FC and PWP, but the effect is more pronounced at FC than at PWP.

Studies showed that water is not equally available to plants over the entire range of available soil water (between FC and PWP). It is observed that yield declines drastically when soil water content falls below a particular point within this range. This point is called **critical soil water level** or **critical soil water tension** for crop yield. In most cases, crops yield optimum when the soil water is maintained from FC to 50% of available soil water, occasionally from FC to 25% (for shallow-rooted leafy vegetables) of available soil water, known as **maximum allowable depletion**.

Soil Water Movement in Saturated Soil

The movement of water through a porous body occurs because of the difference in potential energy between two points. The difference in potential energy creates a driving force responsible for the movement of water from a position of higher potential energy to lower potential energy. The total soil water potential is the sum of the gravitational, pressure, and osmotic potential. The osmotic potential is important for the movement of soil water only when pore size acts as a semipermeable membrane, i.e. when pores will allow water to move, but not the solute molecules. Therefore, for most soils, the sum of pressure (ψ_p) and gravitational potential (ψ_z), which is sometimes called hydraulic head (H), is the driving force for the movement of water in the soil.

$$\psi_t = \psi_z + \psi_p = H$$

The equation, known as Darcy's Law, is the backbone of many steady-state saturated flow descriptions.

$$q = K_s\left(\frac{\Delta H}{L}\right) \text{ or } q \propto \left(\frac{\Delta H}{L}\right)$$

where q is the flow volume per unit cross-sectional area per unit time, called **flux density** (or simply **flux**), thus, flux is proportional to the hydraulic gradient. The proportionality factor K_s is known as the **saturated hydraulic conductivity** of the soil.

In the case of unsaturated soils, the hydraulic conductivity [K(θ)] is a function of moisture content and matric suction (ψ_m), and Darcy's equation can be written as follows:

$$q = K(\psi_m)\frac{\partial H}{\partial x}$$

$$q = K(\theta)\frac{\partial H}{\partial x}$$

The flow equation in a one-dimensional vertical system is:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left[K(\psi_m)\frac{\partial(\psi_m)}{\partial z}\right] + \frac{\partial K(\psi_m)}{\partial z}\Bigg]$$

2.7 SOIL AIR

Soil air means the air present in the soil. Since air and water mutually share the pore spaces of soil, it is not surprising that many physical properties viz texture, structure and porosity affect soil aeration. For the growth and development of plants and soil microbial activity, soil aeration status is as important as the soil water status. Aerobic respiration of plant roots and microorganisms continuously consumes oxygen and evolves carbon dioxide. Therefore, a rapid exchange of these two gases between soil and atmosphere is essential to maintain a favourable aeration status in soil. If the supply of oxygen is not sufficient to meet the demand in soil, organic matter decomposition or in other words, mineralization of plant nutrients and the metabolic activities of plant roots and soil microorganisms are seriously affected, which is ultimately reflected in poor plant growth. For most upland crops, the concentration of oxygen in the soil air must be above 10% (volume basis), compared to 21% in atmospheric air.

The concentration of oxygen in poorly drained soil is always lower than that of well-aerated soil and may drop to less than 5% and even zero. The amount and composition of soil air are governed by the soil water content. In the field rice soil, the concentration of carbon dioxide, methane, and ammonia is higher than that of the atmospheric air. Table 2.2 depicts the comparison in the composition of soil air and atmospheric air.

Plant growth is adversely affected by at least three different ways due to poor aeration: (a) due to poor root growth, (b) absorption of nutrients and water is decreased, and (c) production of some compounds that are toxic to the plant.

2.7.1 Characterization of Soil Aeration Status

There are several ways to characterize the aeration condition of soils. Some of which are discussed here.

Air-filled Porosity: The air-filled porosity is the volume fraction of total pore space not filled with water. Air-filled porosity can be evaluated by an air pycnometer or from a soil moisture characteristic curve.

Redox Potential: The aeration status of soil can be judged by the oxidation/reduction state of the elements. In well-aerated soil, the multivalent elements always remain in their higher valence state, e.g. iron as Fe^{3+} (FeOOH), manganese as Mn^{4+} (MnO_2), nitrogen as N^{5+} (NO_3^-), and sulphur as S^{6+} (SO_4^{2-}). On the other hand, in poorly aerated soil, these elements present in lower valence state, e.g. iron as Fe^{2+} (FeO), manganese as Mn^{2+} (MnO), nitrogen as N^{3+} (NO_2^-), N^{2+} (NO), N^{1+} (N_2O), N^0 (N_2), N^{3-} (NH_4^+) and sulphur as S^{2-} (H_2S). The redox potential value (E_h) of a well-aerated soil may be as high as +0.5 volts. On reduction, the value of E_h may be zero or even may be as low as –0.28 volts.

Composition of Soil Air: The concentration of O_2 and CO_2 in soil air.

Air Permeability: Air permeability refers to the readiness of gaseous exchange in response to the pressure gradient. Air permeability depends on the size and continuity of air-filled pores.

Table 2.2: Comparison in the composition (vol. basis) of soil air and atmospheric air

Component	*Soil air*	*Atmospheric air*
Nitrogen plus argon (%)	79.20	79.00
Oxygen (%)	20.60	20.97
Carbon dioxide (%)	0.20	0.03

Oxygen Diffusion Rate (ODR): This parameter is based on the hypothesis that moisture films present around plant roots and microorganisms limit the O_2 diffusion. The ODR is measured by a platinum electrometer. When a certain electrical potential is applied between a reference and a platinum electrode inserted in the soil, oxygen present on the electrode surface gets reduced resulting in a current flow between the electrodes. This current is proportional to the rate of reduction of oxygen, i.e. the rate of O_2 diffusion to the electrode. The ODR values in soils may vary from 0 to 200 $\mu g\ m^{-2}\ s^{-1}$. The requirement for ODR of several field crops ranges from 38.2 to 65.8 $\mu g\ m^{-2}\ s^{-1}$. The critical ODR values of soil in which the plant roots will not grow are different for different plant species (Table 2.3).

2.8 SOIL TEMPERATURE

Soil temperature is one of the important factors that regulate microbial activity and other processes involved in the growth and development of plants. The decomposition of organic matter, mineralization of plant nutrients from organic form to inorganic form, germination of seeds, and nutrient-and water absorption by plant roots are markedly affected by soil temperature. The soil temperature is directly or indirectly dependent on at least three factors: (i) the net amount of heat absorbed by the soil, (ii) the amount of heat required to bring about a change in soil temperature, and (iii) the energy required for evapotranspiration occurring continuously at or near the soil surface.

Solar radiation is the primary source of energy to warm the soil. But clouds and dust particles intercept incoming solar radiation and absorb (16%), scatter (11%), and reflect (28%) a part of this energy. About 45% of incoming energy reaches the earth’s surface in cloudy humid regions and about 75% in cloud-free arid regions. The global average is 50%. Once the heat is received by the soil, the soil temperature is governed by the thermal characteristics of the soil, viz heat capacity, thermal conductivity, and thermal diffusivity.

2.8.1 Heat Capacity

The heat capacity of a substance is defined as the amount of heat required to change the temperature by one-degree centigrade by heat absorption or release and is equal to its specific heat (s) times its mass (m). Mathematically, it can be represented as follows with the unit of cal/°C

$$C_g = s \times m = \frac{\Delta H}{\Delta t}$$

where C_g is the gravimetric heat capacity (normally expressed in cal/g/°C), ΔH is the amount of heat in calories supplied to the body to increase the temperature Δt (°C).

2.8.2 Thermal Conductivity

The thermal conductivity, κ (cal $cm^{-1}\ s^{-1}\ {}^0C^{-1}$) can be defined as the amount of heat transferred through a unit cross-sectional area in unit time under a unit temperature

Table 2.3: The critical ODR values of some selected crops

Crop	*ODR ($\mu g\ m^{-2}\ s^{-1}$)*
Maize	50
Soybean	47.3
Wheat (irrigated)	43.8
Peas (irrigated)	65.8
Toria	38.2

Source: Sinha and Ghildyal (1982)

gradient between two faces. The thermal conductivity of soil is independent of the temperature gradient, not necessarily temperature. The thermal conductivity of a given soil depends on the relative proportion of solid, liquid, and gaseous phases, size (texture), and arrangement (structure) of the solid particles.

2.8.3 Thermal Diffusivity

It is a measure of the rate of change of temperature in a unit volume of soil due to heat flow under a unit temperature gradient. The thermal diffusivity, D_T (cm^2 s^{-1}) of soil is defined as the ratio of the thermal conductivity, κ (cal cm^{-1} s^{-1} ^{0}C^{-1}), and the heat capacity (cal cm^{-3} ^{0}C^{-1}). Volumetric heat capacity, thermal conductivity, and thermal diffusivity of some soil components are given below (Table 2.4).

$$D_T = \frac{\kappa}{c_v}$$

The thermal conductivity of soil increases with an increase in the moisture content. Initially, when dry soil absorbs moisture the increase in thermal conductivity is small as the major portion of the void space remains filled up with air, a bad conductor of heat. With the gradual addition of water up to a limit ($\cong$ 1 bar suction), the thermal conductivity increases due to increased thermal contact between the particles and forming a continuous path for heat flow. Thereafter, a further increase in water content does not make any significant change in the thermal conductivity.

In general, the thermal conductivity of soil decreases with the decrease in particle size. Thermal conductivity of coarse to fine textured soil follows the order of sand > loam > clay > peat. The decrease in thermal conductivity of soils with the fineness of texture is associated with the reduction of the area of contact between the particles.

Thermal conductivity of soil is observed to increase with the soil compaction. Compaction is associated with a reduction of porosity in dry soil and thus increases thermal contact between particles to facilitate better heat flow. In moist soils, compaction improves thermal conductivity for two reasons: (i) compaction increases the degree of saturation and bridges the gap between particles by water (water is a better heat conductor than air), and (ii) compaction increases the thermal contact between particles as in dry soil.

Organic matter is highly porous and eventually has a lesser thermal conductivity than the soil mineral particles. Again, organic matter promotes granulation (loose, porous soil structure). Thus, an increase in organic matter content reduces the thermal conductivity of soil.

Table 2.4: Thermal properties of soil constituents at 20 ^{0}C and 1atm (average value)

Soil constituent	***Density (g cm^{-3})***	***Specific heat s(cal g^{-1} ^{0}C^{-1})***	***Volumetric heat capacity Cv (cal cm^{-3} ^{0}C^{-1})***	***Thermal conductivity κ (10^{-3} × cal cm^{-1} s^{-1} ^{0}C^{-1})***	***Thermal diffusivity D_T (10^{-3} × cm^2 s^{-1})***
Quartz	2.65	0.175	0.46	20.0	43.0
Many soil minerals	2.65	0.175	0.46	7.0	15.0
Organic matter	1.30	0.46	0.60	0.6	1.0
Water	1.00	1.00	1.00	1.42	1.42
Air	0.0012	0.24	0.00029	0.062	0.21

2.9 SOIL PHYSICAL CONDITION AND NUTRIENT AVAILABILITY

2.9.1 Soil Moisture

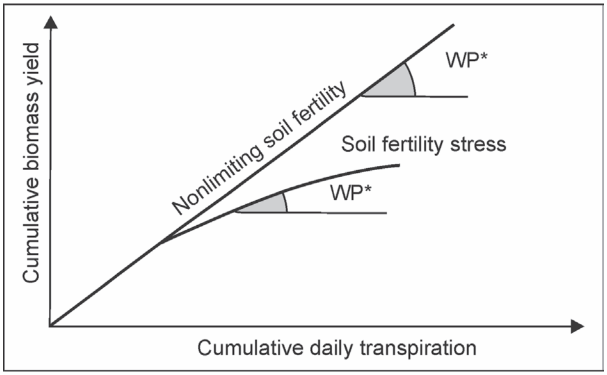
Soil moisture has a close relationship with nutrient availability and plant growth. Better plant growth and nutrient uptake are usually associated with irrigated agriculture. Sometimes, under low moisture conditions, higher salt concentration, particularly with nitrogenous and potassic fertilizers, causes harm to crops. Under moisture-stress conditions, plant growth is limited even with an adequate supply of nutrients, but nutrient content increases, particularly in the case of nitrogen. When plant growth is limited by soil moisture stress, nitrogen tends to accumulate in plant tissues, because the rate of absorption is almost maintained, but the rate of utilization reduces in that situation. However, in any case, if yield increases through fertilizer application, the water use efficiency increases. This may be due to the capability of plants to carry out photosynthesis at a higher rate with sufficient nutrient supply than under fertility stress conditions. Soil fertility stress reduces water productivity as biomass yield increases at a lower rate than non-fertility stress conditions (Fig. 2.4).

Nitrogen: Nitrogen mineralization increases with increasing moisture content from permanent wilting point to field capacity or saturation. Since fertilizer nitrogen is applied to the soil surface, nitrogen uptake decreases upon soil drying. By this time, if the plant cannot uptake excess nitrogen from the soil, the plant may exhibit nitrogen deficiency rather than water scarcity at the onset of drought.

The relationship between soil moisture level and fertilizer application can be summarized as (i) water use efficiency increases with an increase in fertilizer application due to the production of more dry matter, (ii) response of fertilizer application is generally higher under irrigated than unirrigated conditions, and (iii) increasing the levels of fertilizer application enhances the response to frequent irrigation, particularly for vegetative plant growth.

The available literature showed that the root system of crops fertilized with N is nearly four times larger than those of the crops receiving no N, and the roots of fertilized plants below 6-inch soil depth weighed twice as much as of the plants grown in unfertilized soil.

Phosphorus: Generally, P availability to plants is the maximum at optimum soil moisture content. It is essential for the dissolution of P fertilizers, but excessive moisture may reduce



WP* stands for water productivity = Cumulative biomass yield/Cumulative daily transpiration

Fig. 2.4: Relation between soil fertility stress and water productivity

P absorption by plants due to poor aeration or reduced root proliferation. In areas arid regions, the application of P fertilizer close to the seed is more effective than broadcasting.

Potassium: The relationship between soil moisture and K-availability to plants is not fully clear. Some workers observed that the availability of native K increases with an increase in soil moisture content; therefore, a higher response of K-fertilizers is noticed in low rainfall. Potassium fertilization improves the water relations of plants by increasing water content and turgidity of cells through decreasing transpiration. Movement of K ions in soil occurs mainly by diffusion process. At moisture stress conditions, K content in plants is low, as the rate of entry of K decreases to a greater degree than its rate of utilization by plants. However, the effect of soil moisture on K availability depends on soil type. Soils with higher clay content tend to have a much higher ratio of exchangeable Ca/K at higher moisture content.

2.9.2 Soil Aeration

Normal plant growth requires an adequate supply of oxygen to roots. Nutrient uptake by roots is hindered in the absence of sufficient O_2 or increased concentration of CO_2. The nutrient (anions) transport against a concentration gradient requires energy obtained from aerobic respiration in the roots. During active uptake of anions by the mechanism of the cytochrome-cytochrome oxidase system, Fe undergoes cyclic valency change. Hydrogen ions produced during this process probably participate in cation absorption during exchange between cations and H^+. Under poor soil aeration, oxidative degradation of carbohydrates is depressed, and toxic ethyl alcohol is produced by fermentation.

2.9.3 Soil Temperature

It is generally expected that the energy generated in metabolic activity would be spent on nutrient absorption against a concentration gradient. Therefore, factors, like soil temperature, which affect metabolic activity, should have a definite impact on nutrient absorption. The temperature coefficients for the absorption of all ions are not necessarily the same. The temperature coefficient for anion absorption is higher than cation absorption and is lower at higher external concentrations.

The uptake of various nutrients is the maximum at 32ºC/20ºC soil temperature and decreases progressively with deviation to either side. Since the absorption of N and P is well linked with the energy-driven metabolic activity, the deviation is greater in the case of anion absorption (N and P) than cation (K). The highest uptake of nutrients at 32ºC/20ºC soil temperature could be due to greater (i) available soil nutrients, (ii) root growth, (iii) microbial activity, and (iv) mineralization rate of organic matter.

2.9.4 Soil Compaction and Tillage

Instances of soil compaction nowadays are quite common, because of the use of large and heavy machinery in modern agriculture. Compacted soils restrict root penetration and aeration, thereby adversely affecting the ability of plant roots to absorb nutrients. As compaction reduces root growth, plant roots can extract nutrients and moisture from lower volumes of soil. Poor soil aeration and a lower mineralization rate of organic residues adversely affect nutrient absorption in compacted soils, particularly N and P. Reduced N availability in compacted soil results in lower plant N concentration and uptake. When the degree of compactness exceeded 89%, a considerable reduction in N content occurred in spring barley, and to a larger extent in straw than in grain (Fig. 2.5).

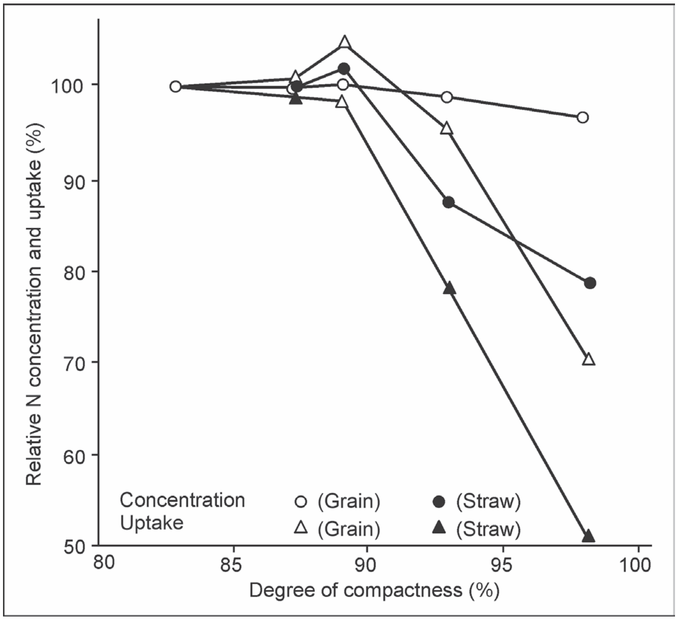


Fig. 2.5: Relative concentration and uptake of N by spring barley as a function of the degree of compaction (*Adopted from J. Lipiec, unpublished data, 1993*)

In recent years, new field management systems have developed, variously called *zero tillage*, *minimum tillage*, or *conservation tillage*. Such systems reduce the number of operations, avoid unnecessary disturbance of topsoil, and generally retain crop residues over the soil surface as mulch. These practices help prevent the reduction of several large pores due to compaction thereby improving soil aeration and infiltration of water.

STUDY QUESTIONS

1. Differences between primary and secondary soil particles.
2. Discuss the binding agents in soil aggregation.
3. Discuss the role of soil porosity in plant growth.
4. What are the different soil moisture constants and their significance in water availability to plants?
5. What are critical soil moisture levels and maximum allowable depletion?
6. What is ODR?
7. Discuss the relation between soil moisture and K fertilizer response.
8. Discuss the behaviour of thermal conductivity at various soil moisture levels.

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