- Freezing and thawing, which causes soil-structure development in cold and temperate climates.

Chemical Processes

Chemical processes of soil formation worth mentioning are:

- The solution of salts;
- The oxidation of organic matter, or, in the formation of acid sulphate soils, of pyrite;
- The reduction of organic matter or iron compounds;
- The formation of clay minerals.

Biological Processes

The processes in which organisms, especially micro-organisms, affect soil formation are:

- Humification (i.e. the decomposition of organic matter and the formation of humus);
- The transformation of nitrogen by ammonification, nitrification, denitrification, and nitrogen fixation;
- Homogenization of the soil resulting from the activities of small animals (e.g. earthworms, termites, moles).

As a result of the soil-forming processes taking place, soil characteristics and properties vary in a vertical direction. Because of the variability in the soil-forming factors (particularly in parent material, landform, and groundwater conditions), soil properties also vary horizontally. These vertical and horizontal variations, which will be treated in more detail in the next section, have great practical implications and are worthwhile studying.

3.3 Vertical and Horizontal Differentiation

3.3.1 Soil Horizons

A soil horizon is defined as a layer of soil or soil material approximately parallel to the land surface and differing from adjacent, genetically-related layers in physical, chemical, and biological properties or characteristics such as colour, structure, texture, consistency, and degree of acidity or alkalinity (SSSA 1987).

Soil horizons that develop as a result of the soil-forming processes are called 'pedogene layers'. When layering is the result of a succession or variation in the parent material, we speak of 'geogene layers'. In young soils with only limited profile development, it is generally easy, at least to the trained eye, to distinguish between pedogene and geogene layering. In old soils with a strong profile development, it is often difficult or impossible to assess whether the layering is due to soil formation only or to a combination of pedogene and geogene layering.

Layers and horizons can have a great impact on drainage conditions because their occurrence determines the flow path that water will take through the soil. Horizons, layers, and their transitions can be identified by differences in texture, structure, consistency, porosity, colour, and various other less easily noticeable differences like calcareousness, salinity, and acidity.

Sometimes, transitions in colour, structure, or texture are conspicuous or distinct, particularly when the soil is dry. More often, however, these transitions are rather diffuse. Though it requires some experience to see these differences, it is unlikely that any important physical transitions are present in case no differences in texture, structure, or porosity can be observed. Chemical differences, or the chemical properties of the soil as such, are rarely directly observable (with the exception of salt crystals). Sometimes, however, chemical differences can be inferred from the shape and size of soil aggregates or from the soil colour. Examples will be given in Section 3.4.2.

3.3.2 The Soil Profile

The soil profile is defined as a vertical section of the soil, through all its horizons, and extending into the parent material (SSSA 1987).

Describing and sampling soil profiles are essential parts of a soil investigation. The soil scientist uses 'master horizons' to describe the vertical sequence of horizons and layers. These are denoted by the capital letters H, O, A, E, B, C, and R. A brief description of these master horizons is as follows:

- H is a wet (anaerobic) organic horizon. Its organic-matter content is more than 30% in clay soils, and more than 20% in sandy soils (Buringh 1979);
- O is a dry (aerobic) organic horizon;
- A is a mineral surface horizon with an accumulation of organic matter;
- E is a mineral horizon from which clay particles, iron oxides, and aluminium oxides have disappeared (also called an eluvial horizon);
- B is a mineral horizon enriched by the translocation of clay particles, organic matter, or iron oxides and aluminium oxides (often called an illuvial horizon);
- C is a mineral horizon of unconsolidated material from which the soil is formed;
- R is a parent rock.

These master horizons can be further divided by suffixes (e.g. 'g' for mottling, 'r' for reduction), or prefixes. For a complete list of definitions and explanations, see the FAO/UNESCO Legend (FAO 1988).

Though soil-science jargon is not very complicated, the non-soil-scientist often has difficulty interpreting the meaning of the horizon codes or is confused by these codes. The major difficulty is how to assess or infer whether a horizon needs designation as, or shows signs of, eluviation (i.e. the leaching of physical and/or chemical soil constituents) or illuviation (i.e. an enrichment due to the accumulation of soil constituents).

The most common horizon sequence of a soil profile is A-B-C. Another horizon sequence found in many highly-developed soils is A-E-B-C (Figure 3.1). For the drainage engineer, B-horizons, and particularly Bt-horizons, are important because such horizons can hinder the flow of water. A Bt-horizon is a texture B-horizon with a higher clay content than the horizon above it.

Horizon sequences can best be observed in specially-dug soil-profile pits at sites

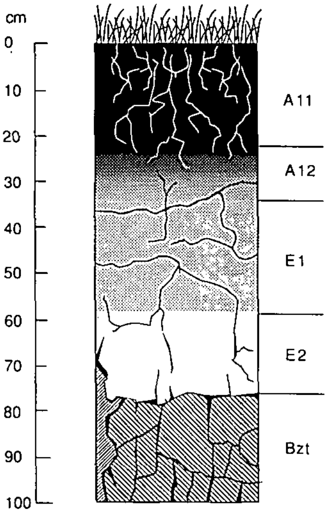

Figure 3.1 Example of a soil profile with an A-E-B horizon sequence (drawn after profile EAK 20 at the International Soil Reference and Information Centre, ISRIC, Wageningen)

that give a representative range of the landscape and vegetation. Alternatively, observations can be made in existing pits or from roadsides or augerings. The various observable characteristics of the soil profile can be described. The data obtained from these observations can be of great help to the drainage or irrigation engineer.

After the profile has been described, samples of each horizon should be taken and sent to the laboratory for chemical, physical, and/or mineralogical analysis.

Apart from the master horizons discussed above, there are also 'diagnostic horizons'. These are used for soil classification, and will be discussed in Section 3.6.

3.3.3 Homogeneity and Heterogeneity

Soil is hardly ever uniform or homogeneous in the vertical direction, and often varies in the horizontal direction as well. For instance, a 'slowly permeable' horizon is hardly ever found at a constant depth. So when using soil maps or making observations, we have to keep in mind that homogeneity in soil characteristics is the exception rather than the rule. Both vertical and horizontal variations are major points of investigation in soil surveys (Section 3.5).

From a pedological point of view, one characteristic that defines a soil is that a certain degree of change has taken place in the profile. A deposit that is uniform from top to bottom cannot, pedologically speaking, be considered a soil because no development of the parent material has taken place. From an agricultural point of view, however, the deposit would be regarded as a homogeneous soil.

A vertical variation in a soil can be partly due to a layered composition of the parent material, but is more commonly the result of profile development (or pedogenesis). Through this development, any vertical homogeneity that might have existed in the parent material disappears. Examples of profile development are the formation and subsequent translocation of organic matter, or the eluviation of clay particles and other compounds along with percolating water.

In soil science and soil surveys, vertical variations and their effects on land use and productivity are the subject of observation and study. They also feature in the keys of various soil-classification systems. Dealing adequately with vertical heterogeneity is not easy, mainly because inferring the quantitative implications for agriculture is so complicated.

Horizontal variations in soil properties are common at any scale, even at less than 1 m. In some cases, the change in colour, salinity, texture, structure, or stoniness/rockiness observable at the soil surface is quite sharp, but, more generally, the transition is gradual. Saline/sodic conditions, in particular, can vary dramatically over short distances; a very saline and sodic profile can change, – within a few metres in the horizontal direction –, to a non-saline, non-sodic profile (e.g. Figure 17.14).

The horizontal variability in soil properties can be studied by various quantitative techniques, which are referred to as 'geostatistics' (Burrough 1986). Geostatistics enable the spatial dependence of data to be determined. This can be used to decide on optimum sampling schemes, to interpolate or extrapolate point observations, and to evaluate how accurately data have been interpolated and extrapolated.

Anisotropy

Anisotropy means that a substance has different physical properties when measured in different directions. In one direction, for example, soil permeability may be higher or lower than in the other direction. Anisotropy can be expected to occur both within a complete soil profile and within soil layers and horizons.

For drainage, it is important to note that the vertical movement of water through the soil is limited by the layer of lowest permeability, whereas the horizontal movement of water is governed by the layer of highest permeability. The vertical movement of water and dissolved salts in the topsoil is determined by water retention and unsaturated hydraulic conductivity (Section 3.4). When considering the general flow path to subsurface drains (Chapter 8), we have to assess whether a soil profile has layers of low permeability, particularly in the topsoil, and layers of high permeability, particularly in the subsoil. The magnitude of the saturated hydraulic conductivity, which is the measure of permeability, will be discussed in Chapter 7, in relation to the shape, size, and orientation of soil grains.

Many structural elements (e.g. prismatic, columnar, and platy structures) are oriented in one direction. This may have its effect on the water-transmitting properties of a soil horizon. In horizons consisting of prisms or columns, there is a similar resistance to vertical and horizontal flow, because the voids around prismatic and columnar elements are interconnected. In surface horizons that contain platy structures, however, the voids mostly occur in the horizontal plane. As a consequence, the horizontal permeability is usually considerably greater than the vertical permeability. In many soils with surface horizons that exhibit surface sealing, the permeability is strongly anisotropic.

Animal activity, particularly when it results in vertical wormholes and the like, can greatly increase the vertical permeability and thus obscure the anisotropy that results from soil horizons or sediment layers having differing permeabilities. Root holes, and cracks in swelling clay soils may have a similar effect.

3.4 Soil Characteristics and Properties

Basic soil characteristics result from the interactions of the soil-forming factors discussed in the previous section. These basic soil characteristics will be discussed in Section 3.4.1. The interactions between them affect a number of physical and chemical properties, which will be discussed in Section 3.4.2.

3.4.1 Basic Soil Characteristics

We distinguish the following basic soil characteristics: texture, mineral composition, physico-chemical characteristics of clay, organic matter.

Soil Texture

The soil consists of primary mineral particles of widely varying sizes. The size distribution of these particles defines the soil's texture. Common names for particle sizes are clay, silt, sand, gravel, stone, and boulder. There are variations in the particle-size limits used by the various disciplines that deal with soils. The USDA/SCS boundary values (Soil Survey Staff 1951, 1975) are listed in Table 3.2A. The major class limits of that USDA/SCS system (i.e. 0.002, 0.050 and 2.0 mm) are widely accepted among soil scientists (see the values used until recently by the FAO in Table 3.2B). Some soil survey organizations (e.g. in the Netherlands), geologists and civil engineers use slightly or completely different boundary values between clay, silt and sand. See for instance the values quoted in Table 3.2C (from the Public Roads Administration in the U.S.) and the new boundary limits recently adopted by FAO (FAO-ISRIC 1990), which apparently are in line with ISO (International Standardization Office) standards (Table 3.2.D).

Texture refers to the particle-size distribution of the 'fine earth' of the soil. These are particles less than 2 mm in diameter (i.e. clay, silt, and sand as defined in Table 3.2A and B).

The textural class of a soil is determined by the relative proportions of sand, silt, and clay in it. The names given to the particular compositions of the sand, silt, and clay fractions vary. Usually these textural classes are presented in a texture triangle. Figure 3.2 shows the textural classification used by many soil survey organizations throughout the world (Soil Survey Staff 1951, 1975; FAO-ISRIC 1990).

The results of particle size distribution from a laboratory analysis can also be presented in the form of a cumulative grain-size curve. Well-graded soils show a good cross-section of particle sizes, ranging from small to large, whereas poorly-graded soils show a uniform particle size or lack medium-sized particles.

Soils are sometimes classified according to their workability. Hence, a coarse-textured soil, in which sand is the dominant fraction, may be referred to as 'light' or 'sandy', and a fine-textured soil, in which clay-particles are the dominant fraction, as 'heavy' or 'clayey'.

Soil texture is important because other properties (e.g. consistency, workability,water retention, permeability, and fertility) are in many cases related to it. If we know the texture of the various layers of soil, we generally have a good idea of the soil's physical properties and its agricultural qualities.

Table 3.2A Particle size limits (Soil Survey Staff 1975)

Soil particle size	Size limits (diameter in mm)		
Clay	< 0.002		
Silt	0.002	-	0.050
Very fine sand	0.050	-	0.100
Fine sand	0.10	-	0.25
Medium sand	0.25	-	0.50
Coarse sand	0.50	-	1.00
Very coarse sand	1.00	-	2.00
Gravel	2.00	-	75
Cobble	75	-	250
Stone or Boulder	> 250		

Table 3.2B Particle size limits (FAO 1977)

Soil particle size	Size limits (diameter in mm)		
Clay	< 0.002		
Silt	0.002	-	0.050
Sand	0.050	-	2
Gravel	2	-	75
Stone	75	-	250
Boulder	> 250		

Table 3.2C Particle size limits of the US Public Roads Administration (quoted by Brady 1990)

Soil particle size	Size limits (diameter in mm)		
Clay	< 0.005		
Silt	0.005	-	0.050
Fine sand	0.050	-	0.25
Coarse sand	0.25	-	2.0
Gravel	2.0	-	?

Table 3.2D Particle size limits (FAO-ISRIC 1990)

Soil particle size	Size limits (diameter in mm)		
Clay	< 0.002		
Fine silt	0.002	-	0.020
Coarse silt	0.020	-	0.063
Very fine sand	0.063	-	0.125
Fine sand	0.125	-	0.20
Medium sand	0.20	-	0.63
Coarse sand	0.63	-	1.25
Very coarse sand	1.25	-	2.00
Fine gravel	2.00	-	6.0
Medium gravel	6	-	20
Coarse gravel	20	-	60
Stones	60	-	200
Boulders	200	-	600
Large boulders	> 600		

Mineral Composition

Two main groups of minerals can be distinguished, depending on particle size:
- Minerals in the sand and silt fraction;
- Minerals in the clay fraction.

The mineral components of the sand and silt fraction are determined by the soil's parent material and its state of weathering. Its composition determines the reserve

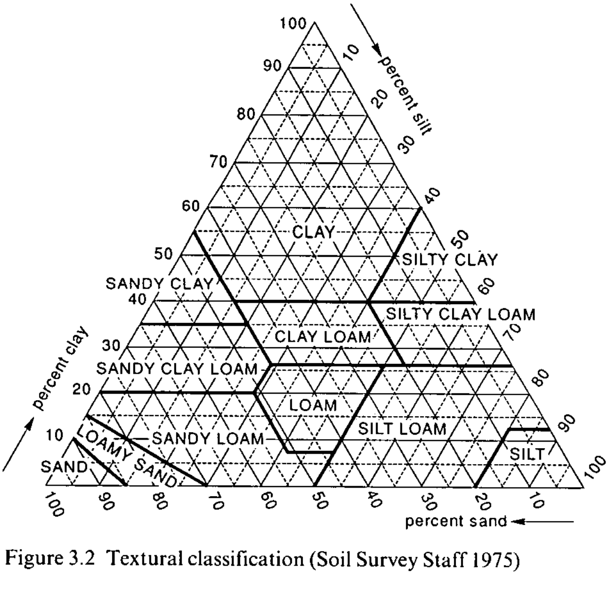

Figure 3.2 Textural classification (Soil Survey Staff 1975)

of minerals available as plant nutrients. The most common component of the sand fraction is silica or quartz which is physically and chemically inert.

The mineral components of the clay fraction consist of crystalline hydrous alumino-silicates. In strongly-weathered tropical soils, we also find crystalline and amorphous iron and aluminium oxides and hydroxides. Hydrous alumino-silicates have a layered structure; they are composed of sheets of silicon oxide and sheets of aluminium hydroxide. A combination of one silicon sheet and one aluminium hydroxide sheet gives a 1:1 type clay mineral. A combination of two silicon sheets, sandwiching an aluminium hydroxide sheet, gives a 2:1 type clay mineral. This layered structure explains why clay minerals occur in plate-shaped crystals. In reality, there are many different clay types that deviate from the ideal 1:1 and 2:1 combinations of silicon oxide and aluminium hydroxide sheets.

The mineral composition of the clay fraction has a direct impact on nutrient availability. Fixation of phosphorus is high in soils with high concentrations of iron and aluminium oxide and hydroxide. Potassium is fixed by clay minerals, the least by tropical kaolinitic clays (see next section) and considerably more by illitic clays (Mitra et al. 1958).

Physico-Chemical Characteristics of Clay

Clays have pronounced physico-chemical properties because of two factors: a large specific surface area, and an electrical charge. The large specific surface (i.e. the surface area per unit mass) results from the platy or fibrous morphology of clay minerals (Table 3.3).

The electrical charge results from a process of isomorphic substitution when the clay minerals were being formed. During that process, some of the silicon and aluminium ions in the crystal structure are replaced by cations of lower valency.

Another factor that creates an electrical charge is the ionization of water on the aluminium sheets into hydroxyl (OH^-) groups. As a consequence, clay particles possess a negative charge at their surface, although some positive charges may occur at the edges of the sheets. This negative surface charge is compensated by the adsorption of positively-charged cations like calcium (Ca^{2+}), magnesium (Mg^{2+}), sodium (Na^+), potassium (K^+), hydrogen (H^+), ammonia (NH^{4+}), and aluminium (Al^{3+}). These cations are present in the so-called 'diffuse double-layer' between clay particles, and their concentration is much higher near the surface of the clay particle than away from it. The adsorbed cations are exchangeable with the cations in the soil solution.

Table 3.3 Specific surface area of various clay minerals

Clay mineral	Specific surface area (m^2/g)
Kaolinite (1:1)	1 - 40
Illite (2:1, non-expanding)	50 - 200
Smectite or montmorillonite (2:1, expanding)	400 - 800

The cation exchange capacity (CEC) refers to this process of mutual replacement (Section 3.4.2).

Thus clay particles are generally platy-shaped and have a high specific surface area. As a result of their chemical composition and spatial arrangement, the 2:1 type clays, such as the montmorillonite (belonging to the smectite group of clay minerals) of subtropical and tropical Vertisols (Section 3.6.5), have substantial electrical charges that bring with them properties like a large CEC, and swelling and shrinking.

The 1:1 types of clays, such as the kaolinite of many tropical clay soils, do not have these electrical charges. These clays have a low CEC and do not swell or shrink.

Many types of clay have properties intermediate between these two extremes. This aspect of clay mineralogy complicates the interpretation of soil-texture data. A soil containing 40% of montmorillonitic clay, for example, behaves quite differently, and also feels finer and heavier, than a soil containing 40% of kaolinitic clay. The latter may feel like a loam and often is called loam (e.g. a Kikuyu red loam which texturally is a clay).

Organic Matter

Organic matter is that part of the soil that consists of organic carbon compounds (i.e. the material derived from the remains of living organisms). When fresh organic matter is incorporated into the soil, part of it is rapidly decomposed by the action of micro-organisms. The residue is called humus, which decomposes slowly and consists of a mixture of brown to black amorphous substances.

Even when present in small amounts, organic matter has a great influence on the physical and chemical properties of soils. Organic matter promotes the stability of soil aggregates, thereby improving the structure of the soil. Chemically, organic matter plays a role in extracting plant nutrients from minerals. The humus component of organic matter increases the CEC of the soil. Moreover, there can be a fixation of nitrogen from the air by micro-organisms, which obtain their energy from decomposed plant tissue.

In some cases, small amounts of organic matter (i.e. of the order of 1%) can have a pronounced effect on soil fertility, but it should be emphasized that a large amount of organic matter does not necessarily make a good soil.

Peat is accumulated organic matter, often to a large degree undecomposed. A combination of a wet climate and poor natural drainage often results in the formation of peat because, under these conditions, the quantity of organic matter produced exceeds the quantity decomposed. By volume, peat soils have an organic-matter content of more than 0.50, muck soils have between 0.50 and 0.20, organic soils between 0.20 and 0.15, and mineral soils less than 0.15 (organic matter as a fraction of dry solids).

Large organic-matter percentages are generally associated with a particular mode of soil formation. When organic matter has accumulated under conditions of poor drainage, the reclamation of such soils often creates problems, such as soil subsidence (Chapter 13), or a very low soil fertility (see Beek et al. 1980).

For a more comprehensive evaluation of the role of organic matter in (tropical) soil fertility, see Sanchez (1976).

3.4.2 Soil Properties

Soil Consistency

The consistency of the soil refers to the effect of the physical forces of 'cohesion' and 'adhesion' within the soil at various water contents. The terminology used ranges from 'loose' to 'extremely hard' in dry soil, from 'loose' to 'extremely firm' in moist soil, and from 'non-sticky, non-plastic' to 'very sticky, very plastic' in wet soil. For more details, see the guidelines for soil-profile description (FAO-ISRIC 1990).

Consistency is related to the type of clay mineral and to the soil chemical status. The consistency is generally lower for coarse-textured soils than for fine-textured, lower for kaolinitic clays than illitic clays, and lower for sodic (see further) than for non-sodic clays. Consistency may be useful in identifying sodicity. Consistency has relevance for soil workability.

In engineering, the classification of soils is often based on texture and plasticity. For this classification, two consistency limits (known as the Atterberg limits) are defined:

- The liquid limit, w_L, which is the minimum water content at which a soil-water mixture changes from a viscous fluid to a plastic solid;
- The plastic limit, w_P, which is somewhat arbitrarily determined in the laboratory as the smallest water content at which soil can be rolled into a 3 mm diameter thread without crumbling.

The plasticity index, PI, equals the liquid limit minus the plastic limit, thereby defining the range of water contents at which the soil behaves like a plastic solid. The plasticity index has relevance for the soil's bearing capacity.

Soil Structure

The structure of a soil is the binding together of soil particles into aggregates or peds, which are separated from each other by cracks. Many wet soils, and also all sandy soils, lack soil cracks and are thus structureless. Structural elements (i.e. the aggregates or peds) can vary in size from a few millimetres to tens of centimetres. The peds can be smooth-edged or sharp-edged, granular, blocky, platy, prismatic, or columnar

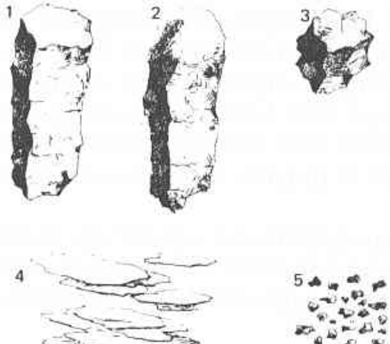

Figure 3.3 Drawings illustrating some types of structural elements (at different scales): 1) prismatic, 2) columnar, 3) blocky, 4) platy, and, 5) granular (Soil Survey Staff 1975)

(Figure 3.3). As the clay content increases, the edges of aggregates get sharper and more prismatic. Structure is related to texture and consistency. It has a positive effect on aeration and permeability.

Particular structures or structural sequences are characteristic of certain soil types (Section 3.6.5).

Soil Colour

Soil colours are primarily due to coatings on the surface of soil particles. The colours can be described according to the Munsell Colour Chart or something similar.

Colour depends on the nature of the parent material from which the soil was formed, on the drainage conditions, and on the prevailing soil temperatures.

Colour variation, whether between soils or within a soil profile, is a useful guide in making a first assessment of general soil conditions. Well-drained and poorly-drained soils have different colours: well-drained soils are redder or browner than poorly-drained soils, which, under similar climatic conditions, are greyer. Black usually indicates organic matter, except in dark-coloured montmorillonite, which generally has a low organic-matter content. In tropical or subtropical regions, red indicates well-drained soils. Yellow may indicate sand or sandy soil in any climate, or, in semi-arid or arid areas, that little soil development has taken place.

Mottling (i.e. the presence of brownish/rusty and bluish/greyish spots) is characteristic of soils in which the watertable fluctuates. Brownish spots occur in the higher parts of layers that are alternately oxidized and reduced as a result of wetting and drying. Bluish/greyish spots occur in the lower part of the groundwater fluctuation zone. In the permanently wet zone, the mottles disappear and uniform grey colours prevail. These bluish grey colours result from the reduction of iron; the reduced conditions are referred to as 'gley'.

Mottles are quite stable and often remain even when the drainage conditions have been improved. Hence, care has to be exercised in interpreting mottles.

Soil Phases, Definitions

The soil consists of three phases: the solid phase, the liquid phase, and the gaseous phase. Methods of quantifying the distribution of the soil phases will be discussed in Chapter 11.

The definitions of some physical soil properties are summarized below.

A volume of soil, V, contains a volume of solids, V_s, a volume of water, V_w, and a volume of air, V_a.

$$V = V_s + V_w + V_a \tag{3.1}$$

The liquid and gaseous phases together form the pore space of the soil, which is occupied by the volume of voids, V_v.

$$V_v = V_w + V_a \tag{3.2}$$

If the voids are completely filled with water, the soil is said to be saturated. The porosity, ε, is defined as the volume of voids as a fraction of the volume of soil.

$$\varepsilon = V_v / V \tag{3.3}$$

A sample of soil can also be divided into mass fractions. Thus, a mass of soil, m, consists of a mass of solids, m_s, a mass of water, m_w, and a mass of air, m_a. In general, m_a can be neglected, so we can write

$$m = m_s + m_w + m_a \approx m_s + m_w \tag{3.4}$$

The wet bulk density, ρ_{wb}, is defined as the mass of soil divided by the volume of the sample.

$$\rho_{wb} = (m_s + m_w) / V \tag{3.5}$$

The dry bulk density or bulk density, ρ_b, is defined as the mass of oven-dried soil, m_s, divided by the volume of the sample.

$$\rho_b = m_s / V \tag{3.6}$$

The soil porosity, ε, can be determined from the density of solids, ρ_s (m_s/V_s, i.e. the mass of solids per unit of volume of solids), and the dry bulk density, ρ_b, according to the equation

$$\varepsilon = (1 - \rho_b / \rho_s) \tag{3.7}$$

The density of dry solids of mineral soils usually varies between 2500 and 2800 kg/m^3. A fair average is 2660 kg/m^3. The density of soils that are rich in organic matter, is lower.

The soil-water content on a volume basis is defined as

$$\theta = V_w / V \tag{3.8}$$

and on a mass basis as

$$w = m_w / m_s \tag{3.9}$$

Coarse and medium-textured mineral soils have dry bulk densities generally varying between 1300 and 1700 kg/m^3. The porosity may thus range from 0.36 to 0.51. In fine-textured soils the dry bulk density is generally somewhat lower than in coarse/medium-textured soils and can be as low as 1100 kg/m^3 (with a porosity as high as $1 - 1100/2660 = 0.60$) in young clay soils. Peat soils have bulk densities lower than that of water (i.e. less than 1000 kg/m^3). Since ρ_s is lower in peat than in mineral soils, the porosity of a peat soil exceeds the range of values indicated for mineral soils.

The bulk density and the porosity cannot be directly related to other soil properties (e.g. permeability). There is the seeming paradox that many soils with a high bulk density and a low porosity have a high permeability, while other soils with a low bulk density and high porosity have a low permeability. This is related to the pore-size distribution.

Pore-Size Distribution

Big pores retain little or no water, but are very effective in conducting water under saturated or nearly saturated conditions (flooding, ponding rain). The opposite is true for small pores, which have a function in water retention, and conduct water slowly. Part of the water in these pores can be taken up by plant roots. When considering the size and the function of the pores, we make a distinction between micro-pores (3 to 30 μm diameter), meso-pores (30 to 100 μm diameter), and macro-pores (> 100 μm diameter).

A soil with an optimum pore-size distribution for plant growth has sufficient micro- and meso-pores to retain water, and sufficient macro-pores to evacuate excess water. Macro-pores are mainly created by soil fauna (earthworms etc.), so increasing the populations of soil fauna is one way of improving the drainage conditions and aerating soils.

The pore-size distribution, which strongly influences a soil's water-retaining and water-transmitting properties, is of great importance for the physical processes of transport in soil. It can be qualitatively assessed by visual observation in soil profiles. Macro-pores are visible to the naked eye; meso-pores are visible at a magnification of 10; micro-pores are not visible, but their presence can sometimes be deduced from the faces of the aggregates, a rough surface indicating the presence of many micro-pores. No field methods are available for quantitative assessments of the pore distribution.

Soil-Water Retention

In a soil, the solid phase usually controls the form or spatial distribution of the liquid and the gas phases. The solid phase is therefore called the 'soil matrix' (Figure 3.4).

Over most of the wetness range in which plant roots normally function, all properties of soil-water retention and transmission are determined by forces associated with the soil matrix. Interactions between the soil matrix and the water are basically due to the forces of adhesion and cohesion. For more details, see Chapter 11.

The availability of soil water is related to its energy status, which is referred to as the 'water potential'. The water potential is governed by the matric forces and by the force of gravity. Other factors may also affect the water potential: the osmotic pressure of dissolved salts, the external gas pressure, and the pressure arising from the swelling of clay. For our purposes, we define the water potential as the sum of the matric potential and the gravitational potential.

The existence of the matric potential can be demonstrated by means of a tensiometer placed in the soil (Figure 3.5). Provided the soil is not saturated, water will move from the porous cup of the tensiometer into the soil. At equilibrium, a negative pressure is measured on the tensiometer.

If we express the soil-water potential per unit weight, we obtain the hydraulic head,

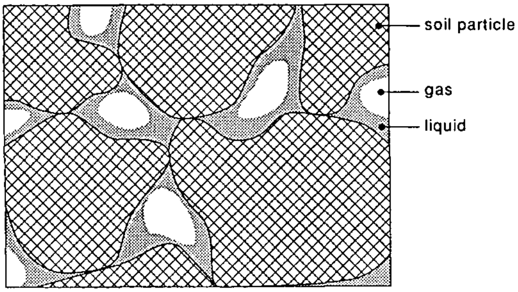

Figure 3.4 Cross-section of soil; soil particles forming soil pores, partly filled with liquid and gas

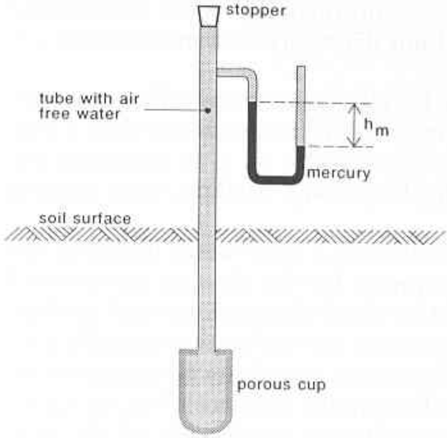

Figure 3.5 A tensiometer

h, being the sum of the pressure head, $p/\rho g$, and the elevation head, z. The pressure head of water in the unsaturated zone is commonly called the 'matric head', h_m. Thus we can write

$$h = h_m + z \tag{3.10}$$

The elevation head depends on the difference between the level of the point where we define the energy status of the water, and a reference level. Usually, the watertable is taken as the reference level.

Above the watertable, the matric head has a negative value because work is needed to extract soil water from the soil pores against the action of the matric forces. This requires a negative pressure or suction. The matric forces decrease sharply when the radius of the pores increases.

The matric head is a function of the soil-water content. At the level of the watertable, the matric head $h_m = 0$, and in oven-dried soil $h_m = -10^7$ cm (= -10^5 m). The graphic presentation of the relation between the matric head and the volumetric soil-water content is called a 'soil-water retention curve' (Figure 3.6). The matric head is conveniently expressed as pF, according to

$$pF = \log |h_m| \tag{3.11}$$

in which h_m is the numerical value of the matric head in cm and pF a number between 0 and 7.

Imagine that free drainage occurs in a soil that has become saturated after a heavy rainstorm. If the soil has large pores in which the matric forces are small, these pores will release water by gravity flow. After this water is released, the soil is at 'field capacity', corresponding with a volumetric soil-water content at a matric head somewhere between –100 and –200 cm ($2.0 < pF < 2.3$). The soil-water content will further decrease by crop transpiration and evaporation at the soil surface. The remaining soil water redistributes by flow through capillaries and flow along the walls of empty pores. When the matric head $h_m = -16000$ cm (pF = 4.2), the soil is at 'wilting point', because plant roots cannot extract water from the soil when the matric head falls below this point. The soil water stored between field capacity and wilting point is called the 'available soil water' or the soil's 'water-holding capacity'.

Figure 3.6 shows the soil-water retention curves of three different soils. Usually, pF-curves are measured by the stepwise drying of a wet sample (desorption). When a dry soil sample is wetted (adsorption), a somewhat different pF-curve will be obtained. This effect is due to pore geometry, and is called 'hysteresis' (Chapter 11).

When the watertable is at shallow depth, the matric head at field capacity is less well-defined, because, if the watertable influences the soil-water conditions in the rootzone, free drainage will be prevented.

If the watertable is lowered, a certain amount of water in the unsaturated part of the soil profile will be released by gravity flow. The 'drainable pore space', μ, indicates the ratio between the change in the amount of soil water and the corresponding change in the level of the watertable.

$$\mu = \frac{\text{change in the amount of soil moisture storage}}{\text{change in watertable depth}} \tag{3.12}$$

Note that the drainable pore space is not a constant for the entire soil profile, but depends on the depth of the watertable.

The drainable pore space is equivalent to the 'specific yield', which was defined in Chapter 2. It is also called 'drainable porosity', or 'effective porosity'.

The drainable pore space of a soil can be found by simultaneously measuring watertable fluctuations and drain discharges over a number of weeks or months. Such measurements integrate the effect of spatial variability of other soil properties. The drainable pore space can also be found from the pF-curve, provided this curve is determined on undisturbed soil samples. Methods of determining the drainable pore space will be discussed in Chapter 11.

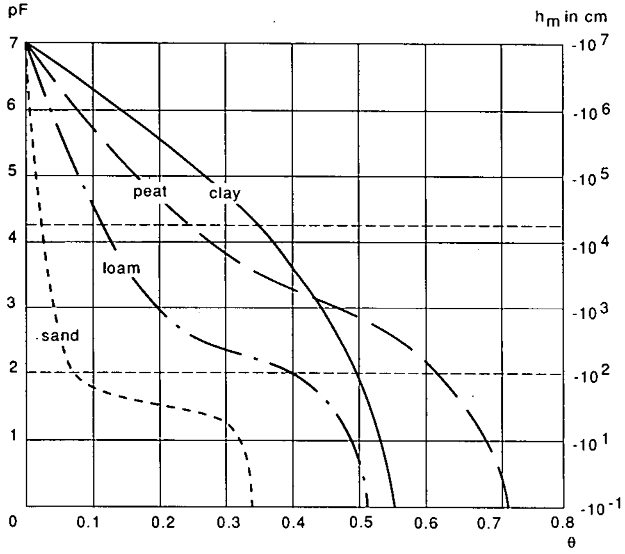

Figure 3.6 Soil-water retention curves for different soil types

Water-Transmitting Properties

Water-transmitting properties of soils can be discerned on the basis of the direction of flow, the position in the soil profile, and the soil-water conditions. The rate of water movement in the soil is governed by the hydraulic head and by the permeability of the soil.

The term 'permeability' has a general meaning and refers to the readiness with which a soil conducts or transmits water. Permeability is expressed by the hydraulic conductivity, which is the proportionality factor in Darcy's Law (Chapter 7). The hydraulic conductivity for saturated flow, K, was defined in Chapter 2. The hydraulic conductivity for unsaturated flow is a function of the matric head, K(h), or the soil-water content, $K(\theta)$ (Chapter 11).

The hydraulic conductivity for unsaturated flow, $K(\theta)$, decreases very rapidly with decreasing soil-water content. One practical consequence is that the flow rates at low soil-water contents are much lower than the potential crop-transpiration rate. In other words, only a part of the available soil water (i.e. the water between field capacity and wilting point) is readily available for plant growth.

'Infiltration' and 'percolation' are processes in which water flows downward at unsaturated or nearly saturated conditions. Infiltration refers to the entry of water into the soil at the surface; percolation refers to the passage of water through the various soil layers. The amount of water percolating through the entire soil profile and recharging the groundwater is called 'deep percolation'.

In small pores, water will rise until the matric forces exerted by the soil particles are in equilibrium with the gravitational force, a phenomenon known as 'capillary rise'. Especially in well-graded soils, capillary rise can reach a height of several metres above the watertable, where water is taken up by plant roots or lost by evaporation at the soil surface. If there is no groundwater recharge, capillary rise causes the watertable to fall until the capillary flow finally stops. If the groundwater is recharged by lateral or vertical inflow (seepage), the capillary flow can continue throughout the season and may transport large amounts of dissolved salts to the rootzone or the soil surface. These accumulated salts can only be removed by percolation which implies a downward movement of water.

Soil Air

Plant roots and most micro-organisms utilize oxygen (O_2) from the soil air and release or respire carbon dioxide (CO_2). A continuous supply of oxygen is needed for a sustained respiration process. An insufficient supply will limit plant growth.

When soil air and atmospheric air are compared, the nitrogen (N_2) content in both is about the same (79%), but the carbon-dioxide content in the soil is higher than in the atmosphere, and the oxygen content in the soil is lower than in the atmosphere. Under conditions of waterlogging, the carbon-dioxide content rises and oxygen may be in short supply.

The interchange of gases between soil and atmosphere takes place by 'diffusion' and by 'mass flow'. Mass flow plays a role when the pressure between the soil air and the atmospheric air differs. These pressure differences may be induced by soil-water flow. With diffusion, gases move in response to their own partial pressure differences. The rate of diffusion is determined by the porosity, and especially by the continuity of the pores. Pore size has little effect on the rate of diffusion, but compacted layers in the topsoil or crusts at the soil surface have a strong adverse effect on soil aeration.

Soil Temperature

Soil temperature is an important growth factor. Below a temperature of 10 °C, microbiological activity is restricted; above 10 °C, the activity increases greatly. Germination depends on the temperature of the topsoil. A low subsoil temperature limits root growth in early spring.

Soil temperature depends, among other factors, on the 'specific heat capacity' of the soil. The specific heat capacity of a dry mineral soil is only one-fifth of the specific heat capacity of water. This large difference explains why wet soils do not warm up as quickly as dry soils. In temperate and mediterranean climates, poorly-drained soils often have soil temperatures 5 °C below the temperature of well-drained soils.

Soil Depth

The term 'soil depth' refers to the rootable depth of the soil. The depth to which plant roots can penetrate into the soil and obtain water and minerals is of great importance for plant growth. When only a very thin soil layer is available for rooting, most plants will experience a deficiency in water and nutrients.

Root penetration is hampered, among other causes, by permanent wetness, by layers of contrasting texture, and, in shallow soils, by cemented or rocky layers. Permanent wetness is easily diagnosed and can, under certain conditions, be remedied by drainage. Similarly, contrasting texture is easy to diagnose and sometimes to remedy by (deep) ploughing. The depth of cemented layers and rock is not difficult to establish either. A cemented layer, however, is often fractured, and plant roots can penetrate through and beyond it.

The effect of a cemented layer or any other type of obstruction to root penetration (e.g. extreme acidity, salinity, sodicity or permanent wetness) needs to be carefully assessed. In practice it is often not easy to establish the actual and potential rooting depth but good observation can help to make the relevant and right estimate.

Cation Exchange Capacity

The 'cation exchange capacity' (CEC) of a soil is defined as the amount of cations that can be adsorbed per unit mass (in cmol/kg or meq/100g). The higher the cation exchange capacity, the more the soil solution is buffered against additions of particular cations, because an exchange of cations can occur between the soil solution and the exchange complex. A small CEC means that small amounts of cations (e.g. hydrogen ions from plant roots) have a pronounced effect on the cation balance of the soil solution.

The range in cation exchange capacity for three kinds of clays and organic matter is given in Table 3.4.

Kaolinite has a low CEC and organic matter a very high CEC. Soils that are characterized by kaolinite as the predominant clay mineral and the absence of appreciable amounts of organic matter, have a very low CEC. Such conditions are common in many tropical soils.

Table 3.4 Cation exchange capacity (CEC) of various clay minerals and organic matter (Young 1976)

Component	CEC (meq/100 g)
Kaolinite	3 - 15
Illite	10 - 40
Montmorillonite	100 - 150
Organic matter	100 - 350

Base Saturation

The 'base saturation' refers to that part of the cation exchange capacity which is saturated with basic cations

$$BS = \frac{\gamma_{Ca} + \gamma_{Mg} + \gamma_K + \gamma_{Na}}{CEC} \tag{3.13}$$

where γ_{Ca}, γ_{Mg}, γ_K and γ_{Na} refer to the amounts (in cmol/kg) of the exchangeable calcium, magnesium, potassium and sodium cations. Low values of the base saturation indicate intense leaching.

Salinity

The presence of soluble salts in the soil solution can affect plant growth, depending on the salt concentration and the susceptibility of the plant or crop. Except in cases of very high salinity where salt crystals can be readily seen, the presence of harmful amounts of salt in the soil is generally not observable to the eye. Soil salinity is appraised by measuring the electrical conductivity or salt concentration in soil-water extracts (Chapter 15). Recently, methods have been developed to measure soil salinity direcly in the field (Rhoades et al. 1990).

Some plants, called halophytes, can withstand, or even like, saline soils. So, in many cases, the vegetation can be a useful guide in identifying salinity, and particularly salinity patterns. Salinity is mostly associated with a near-neutral, slightly alkaline, soil reaction, unless appreciable amounts of sodium are present, when soil reaction is pronouncedly alkaline.

Sodicity

Sodicity refers to the presence of sodium (Na) ions on the exchange complex and in the soil solution. When sodium is present, the soil aggregates are unstable and are likely to disperse. This lack of stability can cause open drains to collapse or pipe drains to silt up. Other major effects are a reduction in soil permeability, a disturbance of nutrient equilibrium, and toxicity to plants. The physical behaviour of sodic soils will be discussed in Chapter 15. Sodicity, usually expressed by the 'exchangeable sodium percentage' (ESP) and/or the 'sodium adsorption ratio' (SAR), is assessed in the laboratory. The slaking of soil aggregates when wetted can indicate sodicity, and, as remarked earlier, the presence of a columnar structure points to high sodicity.

Sodicity is associated with an alkaline soil reaction. When the pH of the soil solution is higher than 8.2, appreciable amounts of sodium are likely to be present.