1 Composition of soil

1.1 Description and classification of soil

The relatively thin mantle of soil over the land surface of the earth is a porous material of widely varying properties. Its solid phase consists of the inorganic products of weathered rock or transported material together with the organic products of the flora and fauna that inhabit soil. Some of these products are recognizable remnants in the form of stones, sand grains, and leaf litter, but others like clay minerals and humus result from the profound chemical changes that occur in both inorganic and organic material during the process of soil formation. The resulting soil can range in texture from coarse sands to fine clayey materials and it can range in its organic content from a usual amount that is less than 5 per cent by weight to about 80 per cent in peaty soil.

These and other properties can differ greatly from place to place over the earth’s surface and from top to bottom through the succession of horizons or layers that constitute the soil profile. Hence, it is useful to be able to recognize some order in the occurrence of soil materials in nature. The profile, extending from the soil surface to a depth that includes the zone explored by plant roots, forms the basis for studying soil distribution systematically. For a given set of conditions affecting its development, the profile exhibits predictable properties. Water penetrating downwards carries materials in solution or colloidal suspension to lower levels, and so creates horizons of accumulation or depletion that characterize the profile. The more humid the climate the greater will be the leaching, so that minerals including chlorides, sulphates, and carbonates that can accumulate in arid areas (as in Fig. 1.1) are removed from the profile in drainage waters under wetter conditions. Similarly, the type of parent material can leave its stamp so that, for example, sandy soils are more likely to develop from sandstone and granite, which contain abundant quartz of sand size, than from shale and basalt. Furthermore, the type of vegetation affects the organic matter profile. The extensive grasslands of eastern Europe
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Fig. 1.1. A soil profile in a semi-arid locality (Renmark, Australia). The soil surface is at the second rung down the ladder, a horizon of carbonate accumulation commences at the fourth, and one of sulphate (gypsum) commences between the fifth and sixth rungs.

and North America have, for example, allowed a deep horizon of dark soil to be formed as a consequence of the ramification and decay of fine roots. In contrast to these so-called chernozems or black earths, those soils developed under forest have their organic matter added mainly at the surface as leaf litter, so that it occurs principally on relatively shallow surface horizons in the brown soils and podzols of forested regions. Also, since soils can form to greater depth on flat lands than on slopes subject to erosion, topography influences soil development. Furthermore, profiles, as they appear today, may have been thousands of years in developing to their state of maturity or they may, as in a drained swamp or an eroding hillside, be undergoing relatively rapid changes under the influence of a recent change in environment. Hence time, parent material, climate, vegetation, and topography all have a part in determining the properties of soil.

Russian scientists demonstrated in the nineteenth century that broad zonal groups of soils occurred on a continental scale in patterns that conformed geographically with climate, vegetation, and other environmental factors. The recognition of these great soil groups gave cohesion and impetus to the study of soils and their classification. However, this genetic approach to soil classification has practical limitations because soil origins are often imperfectly understood. In the early soil surveys undertaken in the United States of America at the level of detail of the farm, soils were identified by characteristic properties of the profile rather than by uncertain genetic criteria and this is current practice throughout the world in detailed surveys. The gaps in classi-
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3

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The individual particles that make up the solid framework of soil can be divided arbitrarily into two classes according to whether or not they are smaller than 2 μm in diameter. The clay fraction (<2 μm) includes chemically and physically reactive clay minerals that have formed as secondary products from the weathering of rocks or have come directly from a parent material of transported deposits. The nonclay fraction consists of inert mineral and rock fragments and sometimes secondary concretions, and it can be further divided into silt, sand, and gravel with size limits that vary between the different scales in current use. Three of these scales are set out logarithmically in Fig. 1.2b. Results of particle size analyses can be presented as cumulative curves as in Fig. 1.2a. From these, the amount in the size fractions of any of the scales can be read between appropriate size limits. More usually, results are presented directly in size distribution tables showing the percentage of soil in each frac-
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Fig. 1.2. Particle size analysis. (a) Cumulative curves for three soils with textures of (1) clay, (2) silty loam, and (3) sand. (b) Limiting sizes of the fractions in systems of the International Society of Soil Science (ISSS), United States Department of Agriculture (USDA), Massachusetts Institute of Technology (MIT), and British Standards Institution (BSI). (f. = fine, m. = medium, co. = coarse, v. = very.)

In soil survey reports, percentages are based on oven-dry soil passing a 2 mm sieve. Gravel and stones larger than 2 mm are reported separately if present. The analysis includes a pretreatment of the soil to ensure that its particles are separated so as to behave as individuals rather than clusters. The sizes are reported as diameters but, since the particles are of many shapes, they are to be regarded as effective diameters based on sizes of sieve openings in the case of the large particles and on settling velocities in the case of the smaller.

The size distributions of the three soils in Fig. 1.2 are plotted as points on a triangular diagram in Fig. 1.3a, where the apices represent 100 per cent sand, silt, and clay fractions, respectively. The limiting size between silt and sand fractions is here 50 μm (US Department of Agriculture system), but other triangular diagrams are also available for the systems with limits of 20 μm (Marshall, 1947) and 60 μm (Hodgson, 1974). Areas marked off on these diagrams show the range of composition of soils typical of each of the texture classes used for describing soils in the field. Notice that the words 'clay',...
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‘silt’, and ‘sand’ are used both for the size fractions (where, for example, ‘sand’ means sand size) and for the texture names where ‘sand’ means soil of sandy texture. Texture is estimated in the field by working the moistened soil between the fingers to sense the coarseness or fineness of the nonclay material and the plasticity and strength imparted by clay. This provides useful qualitative information in the preliminary investigation of soil problems, in advisory services, and in soil surveys.

Particle size analysis also can be represented by plotting an average size of the nonclay material against the clay content. The resulting diagram has the advantage over triangular diagrams of being independent of all arbitrary size divisions except that at 2 μm. Texture classes have been marked on such a diagram in Fig. 1.3b using symbols for five grades of clay content and three of the size of nonclay material. As in the triangular diagrams, these classes constitute the central part of the areas where soils of each field texture description (shown in brackets) are found to cluster. The lines that separate the grades of clay content are not vertical because soils of a given clay content feel less clayey as the average size of their nonclay material decreases (Marshall, 1947). This observation has influenced the design of triangular texture diagrams so as to allow a higher average clay content in a silty class than in the corresponding sandy class.

An upper limit for the clay fraction of 2 μm is a widely accepted but arbitrary one. It is possible to eliminate this from texture diagrams by plotting the geometric standard deviation (representing the spread of the distribution) against the geometric mean diameter. This has been done by Shirazi and Boersma (1984) in their adaptation of the texture triangle.

A texture diagram serves as a reference standard for field descriptions and further, in a number of national soil surveys, it is used for purposes of classification to define texture explicitly from the particle size analysis of soil samples. However, it needs to be recognized that when the particle size distribution of a sample is plotted on a texture diagram the texture class into which the point falls will not necessarily correspond to that given by field description. When the descriptions and analyses of large numbers of samples were compared by Marshall (1947) and by Foss, Wright, and Coles (1975) a large scatter of points was found for each texture and only about half fell within the prescribed area of the diagram. Discrepancies arise not only because of the subjective nature of the field examination but because clay minerals, exchangeable cations, organic matter, and cementing agents can strongly influence the field estimate of texture. Field descriptions, made by manipulating moist soil in the hand, reflect soil behaviour as well as particle size. They thus have a significance of their own and correspond with particle size analysis in an average way rather than on a one-to-one basis.
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(a)

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(b)

Fig. 1.3. Relation between texture and particle size distribution. (a) Texture triangle of the United States Department of Agriculture (Soil Survey Staff, 1975). (b) Texture diagram in which particles larger than 2 \( \mu \text{m} \) are represented by an average size (Marshall, 1947). (F = fine, S = sand(y), Si = silt(y), C = clay(ey), L = loam(y).) (Plotted points represent the three soils of Fig. 1.2a.)
1.3 Pore space

The minerals of which the particles are composed reflect the nature of the parent material and the degree of weathering that the soil has undergone. The minerals of the parent materials vary greatly in their stability and hence in their occurrence in soil. Quartz (SiO₂) often dominates the nonclay fraction because of its resistance to weathering, and its abundance in certain parent materials such as granite, sandstone, and surface deposits. Among the primary silicate minerals, muscovite mica, and potassium feldspar are more resistant than other felspathic minerals or the ferromagnesian minerals. The secondary minerals include some that are relatively soluble (e.g., gypsum, calcite, dolomite) and others that are not (e.g., oxides and hydroxides of iron, aluminium, and silicon). Obviously, the less resistant minerals are only common in soils not subjected to severe weathering and leaching or formed from parent material capable of providing them in relative abundance. For further information on these aspects of mineralogy of soils the reader is referred to Brewer (1964).

The above minerals occur chiefly in the nonclay fraction but they are not confined to it, because its lower limiting size of 2 μm is arbitrary. The principal minerals of the clay fraction are secondary products of weathering referred to collectively as the clay minerals. These have characteristic crystal structures, as shown by X-ray diffraction patterns, and they occur mainly in the form of flakes, as shown by the electron microscope (see Fig. 1.4). In addition, there may be material present in the clay fraction that is amorphous to X-rays. The particles of the clay fraction, ranging downward in size to about 5 nm, are the seat of phenomena such as swelling, plasticity, and cation exchange that have profound influence on the physical behaviour of soil.

1.3 Pore space

The way soil behaves depends not only on the kinds and sizes of individual particles, but also on how these are arranged and bonded together. Sometimes they occur as a collection of single individual grains (as in sands), but usually they are linked into clusters or aggregates of varying stability. The properties of the constituent particles are masked by this clustering which thus has a marked effect on soil behaviour. Between the particles, arranged singly or in aggregates, there is an intricate system of pore space on which plant roots, micro-organisms, and soil animals depend for the storage and movement of water and air.

These topics will be dealt with more fully later but it will be useful to outline here some broad structural features of the three-phase system of solid particles, soil water, and soil air. Taking $V_s$, $V_l$, $V_g$, and $V_t$ to represent, respectively, the volume of solid, liquid, gas, and the total or bulk volume of
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Fig. 1.4. Electron microscope photograph of crystals of kaolinite. The length of the line represents 1 \( \mu \text{m} \) (K. Norrish, CSIRO, Adelaide).

A quantity of soil and taking \( m_s, m_t, m_g, \) and \( m_t \) to represent the corresponding masses, we define

\[
\begin{align*}
\text{density of solid particles}, & \quad \rho_s = m_s/V_s \quad (1.1) \\
\text{bulk density of soil}, & \quad \rho_b = m_t/V_t \quad (1.2) \\
\text{porosity}, & \quad \epsilon = (V_t + V_g)/V_t. \quad (1.3)
\end{align*}
\]

Writing Equation (1.3) as \( \epsilon = (V_t - V_s)/V_t \) and combining this with Equations (1.1) and (1.2), we obtain

\[
\epsilon = (\rho_s - \rho_b)/\rho_s. \quad (1.4)
\]
1.3 Pore space

Table 1.1. Range of values found in bulk density and associated properties illustrated by particular cases. Particle density is taken as 2.65 Mg m\(^{-3}\)

<table>
<thead>
<tr>
<th>Description</th>
<th>Bulk density, (\rho_b) Mg m(^{-3})</th>
<th>Porosity, (\epsilon)</th>
<th>Void ratio, (\epsilon)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface soil of wet clay</td>
<td>1.12</td>
<td>0.58</td>
<td>1.37</td>
</tr>
<tr>
<td>Surface soil of loam texture</td>
<td>1.28</td>
<td>0.52</td>
<td>1.07</td>
</tr>
<tr>
<td>Spheres of uniform size in open packing</td>
<td>1.39</td>
<td>0.48</td>
<td>0.91</td>
</tr>
<tr>
<td>Subsoil of sandy texture</td>
<td>1.61</td>
<td>0.39</td>
<td>0.65</td>
</tr>
<tr>
<td>Sandy loam soil compacted by heavy traffic</td>
<td>1.90</td>
<td>0.28</td>
<td>0.39</td>
</tr>
<tr>
<td>Spheres of uniform size in closest packing</td>
<td>1.96</td>
<td>0.26</td>
<td>0.35</td>
</tr>
<tr>
<td>Sandstone</td>
<td>2.12</td>
<td>0.20</td>
<td>0.25</td>
</tr>
</tbody>
</table>

In soil undergoing change of volume as a result of compaction, swelling, or shrinking, \(V_s\) in Equation (1.3) is not constant and there are then advantages in using the void ratio, \(\epsilon\), instead of porosity. This is defined as

\[
e = (V_1 + V_g) / V_s.
\]  
(1.5)

The relation between \(\epsilon\) and \(\epsilon\), obtained from Equation (1.3) by dividing the numerator and denominator by \(V_s\), is

\[
e = \epsilon / (1 + e).
\]  
(1.6)

or

\[
e = \epsilon / (1 - \epsilon).
\]  
(1.7)

In the previous equation, \(\epsilon\) and \(\epsilon\) are dimensionless ratios and density has the dimensions ML\(^{-3}\). The SI units for density are kg m\(^{-3}\), but the fractional SI units of g cm\(^{-3}\) are also used (1 g cm\(^{-3}\) = 1 Mg m\(^{-3}\)). In most soils the average density of the particles, \(\rho_a\), is in the range 2.5 to 2.8 Mg m\(^{-3}\), but the density of soil organic matter is much less than this, and that of the ferromagnesian minerals, iron oxides, and other so-called heavy minerals is greater than 3 Mg m\(^{-3}\). Bulk density, \(\rho_b\), increases with the degree of compaction and tends to increase with depth in the profile because of increasing overburden and decreasing disturbance. In a clay soil it decreases with the degree of swelling. Fine textured soils tend to be less dense than sands especially if the size distribution of the sand is such that the gaps between the larger particles are occupied by smaller ones.

Selected values of \(\rho_b\) are given in Table 1.1, which covers the range ordinarily found in soils and from these \(\epsilon\) and \(\epsilon\) are calculated by taking \(\rho_a\) as 2.65 Mg m\(^{-3}\). The values are appropriate to the described soil conditions and to the porous rock (sandstone), but they are not specific to them and serve only
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as representative examples. On the other hand, those for spheres in open (cubical) and closest (rhombohedral) packing given for comparison are specific to the geometry for spheres of any uniform size (Slichter, 1899).

Methods for measuring particle density and bulk density are described by McIntyre and Loveland (1974a) and by Blake and Hartge (1986a, b). The most usual method for bulk density is to cut out a cylindrical core of known volume and find the mass of the dried soil. It can also be measured on a clod whose volume is found by displacement in water after its surface has been sealed with a resin. This is a good method for many purposes, but the bulk density of a clod will be greater than that of the soil at large if there is much pore space between the clods. Bulk density can be measured in situ from the attenuation of gamma rays transmitted through the soil, as discussed in Section 3.1.

1.4 Water content

The porosity consists of a portion $V_g/V_t$ occupied by soil air and another $V_i/V_t$ occupied by soil water. The amount of water in soil is expressed in this and other ways as

- water content, volume fraction, $\theta = V_i/V_t$  \hspace{1cm} (1.8)
- water content, mass basis, $\theta_m = m_w/m_s$  \hspace{1cm} (1.9)
- degree of saturation, $S = V_i/(V_i + V_g)$  \hspace{1cm} (1.10)

Water content is most commonly measured by finding the loss of mass, $m_w$, on drying in an oven at the arbitrary temperature of 105°C to a constant mass, $m_s$, although there are also other methods, as described in Chapter 3. On combining Equations (1.8) and (1.9), the following relation is obtained for converting from the mass basis to the volume fraction which is generally more useful in field studies:

$$\theta = \theta_m \rho_s / \rho_w.$$  \hspace{1cm} (1.11)

In obtaining Equation (1.11), it is assumed that the density of water is unaffected by being adsorbed in soil so that $m_w/V_1$ is equal to $\rho_w$, the density of pure free water. Values of $\rho_w$ at different temperatures are given in Table 1.7 but, for most applications of Equation (1.11), it can be taken with sufficient accuracy to be $10^3$ kg m$^{-3}$. Also note that the volume fraction, $\theta$, is equivalent to a depth fraction representing the ratio of the depth of water to the depth of the soil profile that contains it. This form is used when examining gains and losses of water in the field, because precipitation and evaporation are also expressed as depths of water.

In a swelling soil whose volume is not constant, water content is sometimes