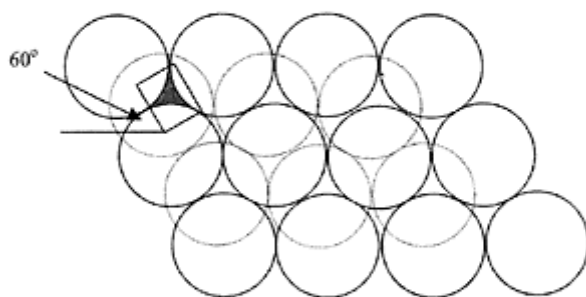
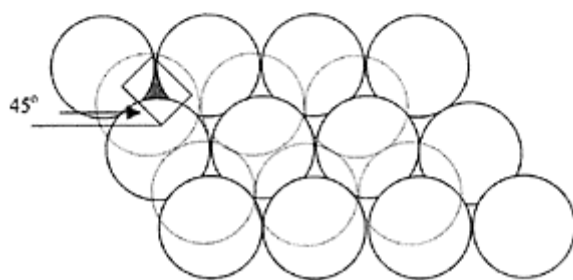


(a) Cubic form



(b) Orthorhombic form



(c) Rhombohedral form

FIGURE 3.19 Different forms of packing of spheres of a uniform size. Within the pore space created by the sphere of radius r in cubic packing, a sphere of radius $r=0.73 r_0$ can be inscribed, but the radius of the interconnected passage is $r=0.41 r_0$. (a) Cubic form; (b) orthorhombic form; (c) rhombohedral form.

Cubic Form. This is the most open form of packing, with the maximum possible porosity of 47.64% or 48%. The porosity can be computed from simple geometric relationships including the volume of the sphere ($\frac{4}{3} \pi R^3$), total volume of the cube with $2R$ sides ($8R^3$), and volume of solids in the cube ($\frac{4}{3} \pi r^3$). Therefore, the pore volume in the cube is computed as follows:

$$\text{Porosity} = \frac{\text{volume of pore space}}{\text{total volume}} = \frac{8R^3 - \frac{4}{3}\pi R^3}{8R^3} = 0.48 \quad (3.49)$$

The pore diameter (d) equals the diagonal of the cube minus the diameter of the sphere or $0.41 D$ where D is the diameter of the sphere. Foster (1932) computed the radius of pores inscribed by uniform spheres of radius r (Figs. 3.19 and 3.20). The radius of the inscribing circle is $0.73 r_0$, but that of the interconnected passage is $0.41 R_0$.

Orthorhombic Configuration. This geometric form involves 3 axes perpendicular to one another. Porosity of such a configuration can be computed as follows:

Total volume of orthorhombal with $2R$ sides = $2R \cdot 2R \cdot 2R \sin 60^\circ \sin 60^\circ = 0.866$

$$\therefore \text{Porosity} = \frac{\text{volume of pore space}}{\text{total volume}} = \frac{4\sqrt{3}R^3 - \frac{4}{3}\pi R^3}{4\sqrt{3}R^3} = 0.40 \quad (3.50)$$

Rhombohedral Configuration. Rhombohedral is a six-sided prism, whose faces form parallelograms.

Total volume of rhombohedral with $2R$ sides = $2R \cdot 2R \cdot 2R \sin 45^\circ \sin 45^\circ = 0.785$

$$\therefore \text{Porosity} = \frac{\text{volume of pore space}}{\text{total volume of rhombohedral}} = \frac{4\sqrt{2}R^3 - \frac{4}{3}\pi R^3}{4\sqrt{2}R^3} = 0.26 \quad (3.51)$$

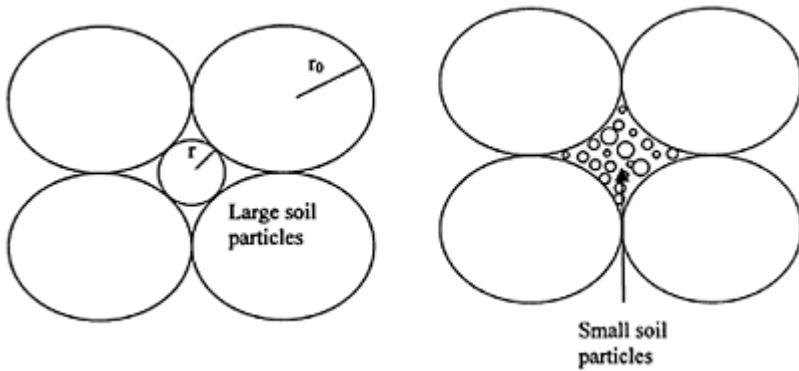


FIGURE 3.20 (a) Open packing; (b) closed packing ($r=0.73 r_0$ in open/cubic packing).

Composite Form. Uniform spheres can also be arranged into composite packing involving cubic and rhombohedral configuration. This situation may happen if soil aggregates or secondary particles were spheres of uniform size. In such a scenario, total porosity of uniform spherical particles within the aggregates in a rhombohedral configuration will be simply the sum of porosity of each configuration.

$$\text{Total porosity} = 0.48 + 0.26(1.00 - 0.48) = 0.62$$

(3.52)

These simple geometric arrangements lead to the following conclusions:

1. For identical form of packing, total porosity is independent of particle size of uniform spheres. However, the maximum pore diameter is proportional to particle size, and hence the permeability varies as a square function of the particle size. This is discussed under Poiseuille's law in Chapter 6.
2. The particles all have the same diameter, the most open packing or cubic form yields a total porosity of 0.48 and the most dense packing or rhombohedral form yields a total porosity of 0.26.
3. If all soil separates or primary particles are aggregated into secondary particles, the total porosity is much greater than when unaggregated.

Close Versus Open Packing

The packing of soil particles is influenced by particle shape and size distribution. For some engineering applications (e.g., dam construction, embankment, foundation, etc.), a high density is required.

Close rather than open packing is normally observed under natural conditions. For this topic readers are referred to the detailed description of packing arrangements by Yong and Warkentin (1966) and Childs (1969). In this regard, the geometry of "close packed" spheres is important to understand. In close packing, the smaller particles are packed within the pore space of larger particles (Fig. 3.20). The close packing is achieved by

arranging the small grain sizes to fill voids created by large particles. Achieving a high density based on close packing necessitates having a material containing a diverse range of particle sizes. The other end of the scale involving open packing is based on a material containing particles of a uniform size. Thus, maximum porosity is achieved with open packing and the least with close packing.

Well-Graded Versus Poorly Graded Material

Packing arrangement of soil material is of relevance to soil compaction and surface seal formation in agricultural soils. It is also of interest to civil engineers concerned with stable foundations. The “well-graded” soil consists mostly of sand and gravel but also contains a small amount of silt and clay to facilitate close packing. “Poorly-graded” soils are those with uniform size fraction, e.g., fine or coarse sand only with little material of other size fractions (Fig. 3.21). Such materials are difficult to manipulate into close packing arrangements, do not compact into a dense mass, and are “poorly graded” soils. Clayey soils, with high swell-shrink capacity and ability to adsorb a large volume of water, are also poor-grade material for construction purposes.

3.2 ORGANIC COMPONENTS

Organic solids form only a small fraction of the total solids (about 5% in surface horizon of many humid-region soils) but play an important role in numerous important soil processes that determine soil quality, its productivity, and environment moderation capacity. Soil organic matter is a complex mixture of living and dead substances of plants and animal origin. Remains of dead plants and animals may be partially or fully decomposed into humic and biochemical substances. There are two principal types of humic substances: (i) insoluble humic acids, and (ii) alkali soluble humic acids and fulvic acids. The latter acids often have high molecular weight.

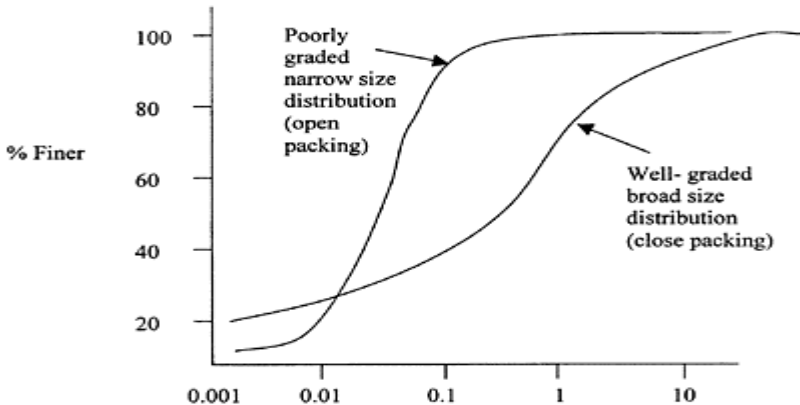


FIGURE 3.21 Particle size distribution for well-graded and poorly graded material.

Humus is dark-colored and amorphous (non-crystalline), and has a low particle density (0.9–1.5 Mg/m³), high surface area, high charge density, high ion exchange capacity, high buffering capacity, and high affinity for water (hygroscopic). In addition to C, humus contains essential plant nutrients including N, P, S, and micronutrients. Because of its high cation exchange capacity (300–1500 cmol/kg), soil organic matter plays an important role in soil fertility management, buffering capacity and ability to filter contaminants from water passing through the soil. It is particularly effective in retaining heavy metals, e.g., Pb, Cd, Cu. Soil organic matter has a high water retention capacity—it can hold 20 times its weight in water. Being highly reactive, humus and other biochemical products are principal ingredients in formation of organomineral complexes, soil aggregates, or secondary particles. Humus forms stable complexes with several elements, e.g., Cu⁺², Mn⁺², Zn⁺², Al⁺³, Fe⁺³. Oxidation or mineralization of soil organic matter can lead to decline in soil structure, and emissions of radiatively active gases into the atmosphere, e.g., CO₂, CH₄, CO, NO, and NO₂.

Depending upon the composition, soil organic matter is classified into several pools. Four principal categories of these pools along with their mean residence time are described in Table 3.15. The easily decomposable fraction is called the “labile or active” pool. The fraction with a long mean residence time is called “recalcitrant or passive” pool. The passive pool may have mean residence time of centuries to millennia. The active fraction has a strong influence on elemental cycling (N, P, S,

TABLE 3.15 Different Pools of Soil Organic Matter

Pool	Constituents	Mean residence time (years)
Labile pool		
(i) Metabolic litter	Plant and animal residues, cellulose	<0.5
(ii) Structural litter	Plant residues, lignin, polyphenol	0.5–2
Active labile pool	Microbial biomass, simple carbohydrates, enzymes	0.2–1.5
Intermediate pool	Particulate organic matter	2–50
Recalcitrant pool	Humic and fulvic acids, organo-mineral complexes	500–2000

Turnover time is calculated by dividing the total pool by flux. For example, if the total soil C pool is 100 Mg and the flux is 50 Mg/ha/yr, then the mean residence time (MRT) is 100/50=2 yrs.

Source: Modified from Parton et al., 1987; Jenkinson and Raynor, 1977; Jenkinson, 1990; Woormer et al., 1994.

Ca, Mg), and on activity of soil fauna and flora. The passive pool influences stability of soil structure through formation of organomineral complexes.

Laboratory determination of soil organic carbon (SOC) is based on methods involving one of the three following principles:

1. Wet oxidation of SOC in acid dichromate solution (Walkley and Black, 1934).
2. Wet oxidation of SOC in acid dichromate solution and measurement of CO₂ evolved (Allison, 1960).
3. Dry combustion of SOC with or without measurement of CO₂ evolved (McKeague, 1976).

Based on these three principles, there is a wide range of methods available for determination of SOC concentration (Nelson and Sommers, 1982; Tiessen and Moir, 1993; Lal et al., 2001). Results obtained are technique dependent, and may vary widely among methods. There is an urgent need to improve upon and standardize the methods of determination of SOC content (Lal et al., 2001).

The soil organic matter pool has a strong impact on the global carbon cycle, and on the atmospheric pool of carbon, especially with regard to the concentration of CO₂. Therefore, assessment of SOC pool, with regards to land use change and soil management, is very important.

Example 3.2

Compute the rate of change in SOC pool upon conversion from natural to agricultural ecosystem if the SOC concentration in 0 to 50 cm depth of a forested soil changed from 2.5% with a bulk density of 0.9 Mg/m³ to 1.2% with a bulk density of 1.2 Mg/m³ over a 10-year period.

Solution

$$\begin{aligned}\text{SOC pool in a forest ecosystem} &= 10^4 \frac{\text{m}^2}{\text{ha}} \times 0.5 \text{ m} \times 0.9 \frac{\text{Mg}}{\text{m}^3} \times \frac{2.5}{10^2} \\ &= 112.5 \text{ Mg/ha}\end{aligned}$$

$$\begin{aligned}\text{SOC pool in an agricultural ecosystem} &= 10^4 \frac{\text{m}^2}{\text{ha}} \times 0.5 \text{ m} \times 1.2 \frac{\text{Mg}}{\text{m}^3} \times \frac{1.2}{10^2} \\ &= 72.0 \text{ Mg/ha}\end{aligned}$$

$$\begin{aligned}\text{Rate of change of SOC pool} &= (112.5 \text{ Mg} - 72.0 \text{ Mg})/10 \text{ yrs} \\ &= 0.41 \text{ Mg/ha/yr}\end{aligned}$$

3.3 IMPORTANCE OF SOIL SOLIDS

Knowledge of soil solids is important to sustainable use of soil resources for different soil functions and land uses. Properties and processes relevant to inorganic solids and their effects are outlined in Table 3.16. Soil solids have an important effect on agricultural and industrial/engineering land uses, and environments. Agriculturally, soil solids are important to soil tillage and trafficability, plant available soil water, leaching losses of fertilizers and chemicals, formation of soil structure, swell-shrink properties, and physical

condition of the soil or soil tilth. In terms of engineering and industrial uses, soil solids are important to foundation strength and stability, water sorption properties, and transmission of fluids in relation to waste disposal. Environmental applications of soil solids are those related to water and air qualities, buffering capacity, and ability to filter contaminants.

There are numerous functions of organic components. The organic components moderate soil and environment qualities. The soil quality effects of organic constituents are due to: (i) improved soil structure, (ii) increased water holding capacity, (iii) increased nutrient availability, and (iv) high soil biodiversity. Environmental effects of soil organic matter are attributed to: (i) high buffering capacity, (ii) chelation with heavy

TABLE 3.16 Importance of Soil Solids to Agriculture, Engineering, and Environments

Property	Agriculture	Engineering	Environments
Texture	Soil tillage and draft power, traffic-ability, soil compaction, plant available soil moisture	Foundation stability, sedimentation	Water quality and air quality effects of sediments
Surface area	Chemical sorption and buffering capacity, leaching of fertilizer	Strength and stability of material	Filtration of pollutants, contaminants, and pathogens
Diffused double layer	Soil structure formation, swell-shrink properties	Water sorption, and foundation stability	Transport of chemicals in water
Packing arrangements	Soil compaction, porosity	Strength and stability of engineering structures, transmission of fluids in relation to waste disposal	Filtration of chemicals

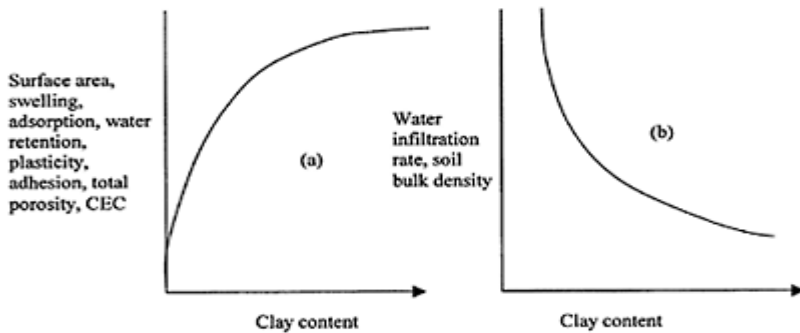


FIGURE 3.22 Relation between clay content and soil properties and processes.

metals and filtration of pollutants and environmental contaminants, and (iii) a large global carbon pool. Soil solids affect numerous properties and processes.

3.3.1 Texture and Soil Processes

Relative proportions of sand, silt and clay affect numerous soil properties. Being the most reactive fraction, increase in clay content increases surface area, swell-shrink capacity, absorption, water retention, plasticity, adhesion, and total porosity (Fig. 3.22a). In contrast, however, increase in clay content decreases water infiltration rate and soil bulk density (Fig. 3.22b). The nature of specific relation depends on other soil parameters (e.g., clay minerals, organic matter content, etc.).

The impact of texture on soil is manifested through its effect on other properties and related processes (Table 3.17). Texture influences soil compaction through its effect on aggregation and porosity, absorption of water and other organic/inorganic compounds by altering surface area, water and nutrient storage through charge properties, transport of solute and gaseous exchange through porosity, etc. In addition to particle size per se, clay minerals also affect surface area, charge density, and in turn, several processes related to these characteristics (Table 3.18).

Textural properties affect agronomic operations and water management. Tillage and traction are strongly influenced by textural properties as well as water content. Soil drainage is strongly influenced by clay content and the nature of clay minerals. There are also numerous engineering applications of textural properties (Table 3.19). Compaction, strength, slope

TABLE 3.17 Soil Properties and Processes Affected by Texture and Inorganic Components

Soil properties	Processes
Bulk density	Compaction, bearing capacity
Surface area	Adsorption, aggregation
Water affinity ^a	Water and nutrient uptake, aeration
Pore size distribution	Transport of solute and solids, leaching, erosion, diffusion
Swelling potential	Cracking, deformation
Plasticity	Moulding, aggregation
Adhesion, cohesion	Formation of soil tilth
Surface charge	Adsorption, absorption, diffusion, chelation
Packing	Compaction

^aNot retention.

TABLE 3.18 Soil Properties and Processes Influenced by Clay Minerals

Property	Processes
Surface area	Absorption, filtration
Charge density	Ion exchange, leaching
Lattice expansion	Swell–shrink capacity
Shape	Plasticity

TABLE 3.19 Engineering Applications

Property	Application
Size distribution	Compaction, strength, trafficability, foundation stability, filtration
Clay content	Absorption, liquid waste disposal
Clay content	Seepage below drain, drainage, ceramic industry
Clay minerals	Slope stability, ceramic

stability, and seepage are strongly influenced by particle size distribution and the nature of clay minerals.

3.3.2 Organic Fraction and Soil Processes

Similar to clay, soil organic matter is also highly reactive. It has high surface area, charge density, and affinity for water. Thus, it has a strong influence on numerous soil properties and processes. The organic fraction influences

TABLE 3.20 Soil Properties and Processes Affected by Soil Organic Component

Soil properties	Processes
Color	Heat absorption, warming
Surface area	Adsorption, aggregation
Charge density	Cation exchange, chelation, aggregation, buffering capacity
Porosity and pore size distribution	Transport of solute and solids, leaching
Bulk density, particle density	Compaction, erosion, bearing capacity
Gaseous composition of soil air	Soil respiration, gaseous emission to the atmosphere
Microbial biomass and activity	Mineralization, aggregation, soil respiration, nutrient immobilization

Plasticity	Moulding, soil tilth formation
Swelling potential	Cracking deformation
Adhesion, cohesion	Soil tilth, soil structure

TABLE 3.21 Agricultural Applications of Soil Texture and Organic Components

Activity	Applications
Tillage	Timing, type, frequency and intensity of tillage
Fertilizer use	Rate, mode, timing, formulation of fertilizer use (precision farming)
Pesticides	Rate and mode of application
Water management	Rate and frequency of irrigation, and intensity of drainage
Accessibility	Timing of farm operations due to warming and trafficability

thermal properties through alteration of soil color, aggregation through charge properties and surface area, nutrient retention through charge density, and soil tilth through aggregation (Table 3.20). Consequently, the organic fraction affects timing and nature of tillage, rate and type of fertilizers to be used, fate of pesticides, and transport of water and pollutants into the soil (Table 3.21). The generic relationship between soil properties and soil organic matter content is shown in Fig. 3.23. Increase in organic fraction increases aggregation, porosity and available water capacity (Fig. 3.23a), and decreases adhesion, cohesion, and shrinkage (Fig. 3.23b). It is because of these improvements in soil characteristics that increase in soil organic content often leads to increase in crop yields (Fig. 3.24). The magnitude of increase in yield, however, depends on soil type and its organic matter content. Such beneficial effects on

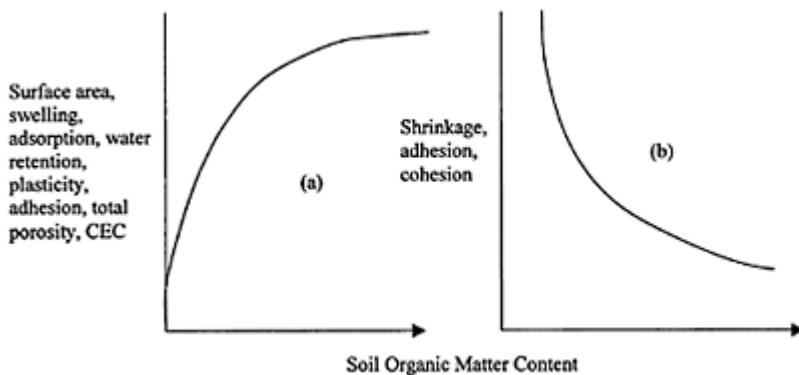


FIGURE 3.23 Relation of soil organic matter content with soil properties.

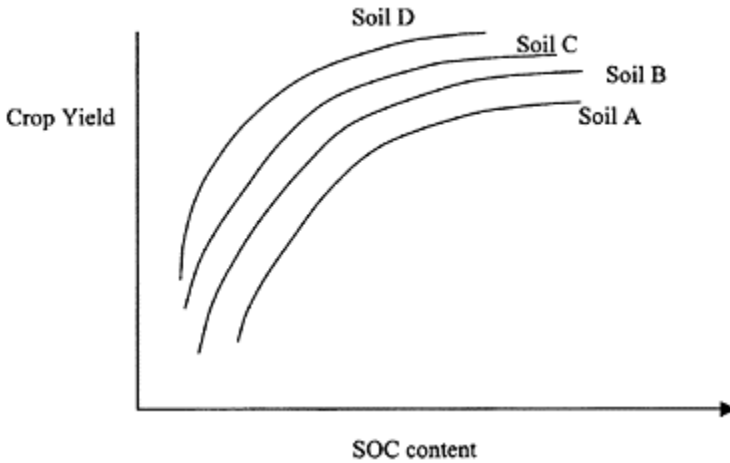


FIGURE 3.24 A generalized relationship between soil organic carbon (SOC) content and agronomic yield.

agronomic yield are especially apparent in subsistence agriculture with low off-farm input.

Beneficial effects of organic fraction on plant growth and yield are also related to improvement in soil quality and decrease in susceptibility to degradative processes. With a strong interaction with texture and clay minerals, the organic fraction affects soil's susceptibility to erosion, compaction, and other degradative processes (Table 3.22). The effects on soil quality are manifested in the overall impact of soil solids (inorganic and organic components) on the environment (Table 3.23). As will be discussed in the chapter on gaseous exchange (Chapter 16), the organic fraction affects flux of several greenhouse gases from soil into the atmosphere.

TABLE 3.22 Soil Degradative Processes Influenced by Inorganic and Organic Components and Clay Minerals

Property	Degradative processes
Texture	Erosion, compaction, leaching, acidification
Soil organic matter content	Acidification, leaching
Clay minerals	Structural decline, crusting

TABLE 3.23 Environmental Applications of Textural Properties and Organic Matter Content

Air quality	Suspended load and particulate matter, smog, soot Gaseous emissions (e.g., CH ₄ , CO ₂ , NO _x , H ₂ S)
Water quality	Suspended load

Important among these are CO₂, CH₄, N₂O, and H₂S, etc. Through its buffering capacity and ability to retain and degrade pollutants, the organic fraction influences water quality.

Sand is the skeleton, clay the flesh, and organic matter the “blood” of the soil.

PROBLEMS

- Calculate the terminal velocity of spherical particles of 2, 0.02, and 0.002 mm diameter in dilute water suspension at 20, 30 and 40°C.
- Calculate specific surface area per unit mass and unit volume of:
 - Spherical quartz particles ($\rho_s=2.65 \text{ Mg/m}^3$) of radii 2 mm, 0.02 mm and 0.002 mm.
 - Plate-shaped particles of length 0.002 mm, width 0.001 mm and thickness 0.00001 mm, and particle density of 2.65 Mg/m^3 .
- Compute specific surface area of the A horizon of a Crosby soil at the Kenny Road Farm with the following characteristics:
 - 60% sand with an average e.c.d. of 0.1 mm, $\rho_s=2.65 \text{ Mg/m}^3$.
 - 30% silt with an average e.c.d. of 10 μm , $\rho_s=2.65 \text{ Mg/m}^3$.
 - 10% clay with platy structure of length=200 nm, width=100 nm and $d=5 \text{ nm}$, $\rho_s=2.8 \text{ Mg/m}^3$.
 - Calculate the relative contribution of each particle size class to the specific surface area,
 - What is the textural classification of this soil? Compare it with that containing 10% sand, 30% silt and 60% clay,
 - What may be possible management problems of these two soils?
- Assume that a soil has a w equal 0.3 and ρ_b equal 1.3 Mg/m^3 . If the soil dries to $w = 0.1$ and shrinks by an amount equal to water loss, calculate ρ_b when $w = 0.1$.
- A quantity of oven dry soil having a particle density of 2.65 Mg/m^3 and weighing 135.8 g is uniformly disturbed in water to form a total volume of 1000 cm³ of suspension. After standing for 3 minutes, 10 cm³ of the suspension removed was 0.437 g. Assuming the temperature of the suspension was 20°C, determine the percentage of particles finer than a specific size fraction.
- Increase in volume of the suspension was 20 cm³ when 50 g of dry soil was mixed in a known volume of water. Calculate particle density if soil bulk density is 1.2 Mg/m^3 .

7. How do charge properties of soil relate to water quality and filtration attributes of soil?
8. What are the agronomic impacts of soil texture and surface area?
9. Briefly describe some applications of Stokes law in natural and altered ecosystems.
10. Describe effects of soil texture on other soil properties and processes relevant to (a) agronomic, (b) engineering, and (c) industrial uses.
11. Using the data in the table below: (a) plot the frequency and summation curves for three soils, and (b) calculate D_{10} , D_{60} , U.C., and gradation coefficients for three soils.

Particle size	% retained		
	A	B	C
4.0	0	20	1
2.4	1	5	2
2.0	2	2	4
1.2	1	6	5
0.6	11	10	8
0.3	7	8	10
0.15	20	12	12
0.075	16	10	10
0.04	5	4	5
0.01	10	10	8
0.002	10	7	5
0.001	17	6	30

12. What are the sources of charge on clay particles?
13. Describe distribution of charge in a fully hydrated clay particle.
14. What are the factors affecting zeta potential? Describe the process of flocculation.
15. (a) Write a brief essay on methods of measuring specific surface area of soil solids,
(b) How much is a net charge on a dried out soil?
16. A farmer in Ohio has shifted from conventional plowing to no till farming. By doing so, SOC concentration in the top 1-m depth is increasing at the rate of 0.01%/yr. Assuming mean soil bulk density of 1.5 Mg/m^3 , calculate the rate of soil carbon sequestration in this 500-hectare farm.
17. Consider a cubic/open packing of spheres of uniform radius of 1 mm. What is the radius of the pore inscribed by four spheres?

APPENDIX 3.1 STANDARD SIEVE SIZES

U.S. Sieve No.	Tyler Mesh No.	Millimeters	Inches
4	4	4.7	0.185
6	6	3.33	0.131
8	8	2.36	0.093
10	9	2.0	0.078
13	10	1.65	0.065
16	14	1.17	0.046
20	20	0.833	0.033
30	28	0.589	0.023
40	35	0.417	0.016
50	48	0.295	0.012
60	60	0.25	0.01
70	65	0.208	0.008
80	70	0.177	0.007
100	100	0.149	0.006
130	150	0.104	0.004
140	170	0.088	0.0035
200	200	0.074	0.0029
400	400	0.038	0.0015

APPENDIX 3.2 COMMON UNITS**Units**

$$1 \text{ dyne} = \text{g} \cdot \text{cm}/\text{s}^2 \quad 1 \text{ dyne}/\text{cm} = \text{g}/\text{s}^2$$

$$1 \text{ Newton} = 1 \text{ kg} \cdot \text{m}/\text{s}^2 = 10^5 \text{ dynes}$$

$$1 \text{ Pascal} = 1 \text{ N}/\text{m}^2 = 10^5 \text{ dynes}/\text{m}^2 = 10 \text{ dynes}/\text{cm}^2 = 1 \text{ kg}/\text{m} \cdot \text{s}^2$$

$$1 \text{ bar} = 10^6 \text{ dynes}/\text{cm}^2$$

$$1 \text{ atm} = 0.101 \text{ MPa}$$

$$1 \text{ J} = 1 \text{ N} \cdot \text{m} = 10^7 \text{ erg}$$

$$1 \text{ erg} = 1 \text{ dyne} \cdot \text{cm}$$

$$1 \text{ cal} = 4.186 \text{ J}$$

$$\text{Poise} = \text{g}/\text{cm} \cdot \text{s}$$

$$1 \text{ W} = 1 \text{ J}/\text{s}$$

$$\text{Radius of H}_2\text{O molecule} = 138 \text{ \AA}$$

1 mole of $\text{H}_2\text{O} = 18 \text{ cm}^3 = 6.02 \times 10^{23}$ molecules
 $\therefore 1 \text{ cm}^3$ of $\text{H}_2\text{O} = 3.34 \times 10^{22}$ molecules

APPENDIX 3.3 INDICES OF PARTICLE SHAPE

Dimensional expression of shape	Index	Formula
2-D	Cailleux's roundness (R) Powers' scale	$R = (2r/a) \times 1,000$ visual comparison chart
3-D	Zingg's classification	based on ratios of b/a and c/b
	Krumbein's sphericity (S) Cailleux's flatness (F)	$S = 3\sqrt{(bc/a^2)}$ $F = ((a+b)/2c) \times 100$

Note: a = long axis; b = intermediate axis; c = short axis; r = minimum radius of curvature at the end of the longest axis of the particle in its plane of maximum projection (measured by comparison with a set of standard concentric semi-circles).

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4

Soil Structure

4.1 DEFINITION AND BASIC CONCEPTS

The arrangement and placement of soil particles determines the response of soil to exogenous stresses such as tillage, traffic, and raindrop impact. This arrangement of soil particles is called “soil structure.” The arrangement is dynamic, complex, and is not very well understood. That is why Jacks (1963) stated that “the union of mineral and organic matter to form the organomineral complexes is a synthesis as vital to the continuance of life as, and less understood than, photosynthesis.” Numerous advances in clay mineralogy, colloidal science, and sedimentology have since led to better understanding of genesis, characterization, and management of soil structure (Yong and Warkentin, 1966; Baver et al., 1972; Revut and Rode, 1981; Larionov, 1982; Burke et al., 1986; Hartge and Stewart, 1995; Carter and Stewart, 1996). Yet, soil structure remains to be the most complex, the least understood, and among the most important soil physical properties.

One of the reasons for the complexity of soil structure is the range of scales it expresses. Structural processes occur at a scale ranging from a few \AA to several cm. Another cause of complexity is the dynamic nature of soil structure. Structural attributes vary in time and space, and the attributes observed at any given time reflect the net effect of numerous interacting factors which may change at any moment. It is truly a moving target. Consequently, it is hard to define soil structure, and the literature is replete with numerous and often confusing terminology, definitions, and approaches. Several terms are used to express easily identifiable structural units including structural form, fabric, aggregate, ped, granule, crumb, tilth, and so on used by different disciplines of soil science.

4.1.1 Different Approaches to Describing Soil Structure

There are at least four related but distinct approaches to describing soil structure. These include pedological, edaphological, engineering, and ecological approaches.

The Pedological Approach

This approach of defining soil structure is based on a mechanistic view with regard to the properties of its components. Therefore, soil structure refers to size, shape, arrangement, and packing of particles into identifiable units called aggregates or peds. In contrast to the synthesis of its components into aggregates, soil structure has also been defined as “the very fragments or clods into which the soil breaks up” (Zakhrov, 1927). In pedological terms, soil structure is a “three-dimensional arrangement of individual mineral grains and organic constituents.”

The Edaphological Approach

This approach is based on its functional attributes with regards to plant growth. Functional attributes of soil structure are those related to pores or voids that govern root growth and development, retention and transmission of water, and gaseous diffusion. It is the soil-pore system that is the most important aspect of soil structure, which includes two types of pores (Fig. 4.1): (i) those within an aggregate are determined by textural characteristics and packing state of elementary particles and are called textural pores or intraaggregate pores, and (ii) those between aggregates and which result from arrangement of structural elements and aggregate characteristics are called interaggregate or structural pores (Stengel, 1990). The most important aspect is the number, dimensions, and continuity of pores between primary and secondary particles. Consequently, soil structure has also been defined as the “assemblage of aggregates (peds) and voids, including voids between and within aggregates” (Thomasson, 1978), or “the

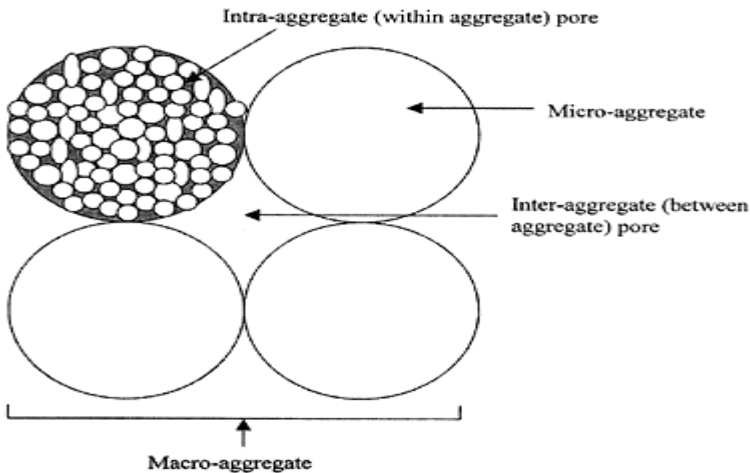


FIGURE 4.1 Interaggregate and intraaggregate pores.

arrangement of solid phase of the soil and of the pore space located between its constituent particles” (Marshall and Holmes, 1979).

The Engineering Approach

It is also appropriate to consider the engineering viewpoint of soil structure. Important among engineering functions of soil structure are bearing capacity, shear strength, slope stability, compressibility, and water permeability (see Chapter 7). With engineering perspective, soil structure is the “strength and stability of aggregates and voids in terms of their compressibility, bearing capacity, and permeability.” Another related term used in engineering is sensitivity, which is the ratio of the strength of an undisturbed soil to that of a soil completely remolded at constant volume. Sensitivity refers to the loss in strength of a soil when its original structure is destroyed by remolding (Wu, 1981).

The Ecological Approach

Perhaps the most complete definition of soil structure is the one that combines pedological and edaphological views and takes a holistic or an ecological approach to soil structure. By so doing, soil structure refers to “size, shape, and strength of aggregates and pores, capacity of pores to retain and transmit fluids and dissolved and suspended materials, and ability to support vigorous root growth and development” (Lal, 1991). In other words, soil structure refers to three aspects: (i) degree of aggregation, their size distribution and stability, (ii) porosity, pore size distribution, shape, tortuosity, continuity, and stability, and (iii) spatial and temporal alteration in aggregates and pores in relation to natural (pedogenesis) and anthropogenic (management) factors. Therefore, in this chapter, aggregation is used to denote pedological (form, shape, size, etc.) and edaphological (functional) aspects of soil structure.

4.1.2 Soil Structure Versus Soil Fabric

Pedologists’ use of the term “fabric” refers to “geometric and spatial arrangement of individual soil particles and voids” (Bullock et al., 1985; Brewer, 1976). In contrast, soil structure includes “the organization of soil constituents into larger aggregates or secondary/compound particles.” Drees (1992) compiled the literature regarding the meaning of these two terms to minimize confusion and inconsistency in their use (Table 4.1). It is important to note that assessment of soil fabric is necessary for a proper evaluation of soil structure (Yong and Warkentin, 1975).

The term “fabric” implies two principal components: the skeleton or the individual mineral grains and the plasma or the soil material that floats

TABLE 4.1 Comparison Between Soil Structure and Soil Fabric

Soil structure	Soil fabric	Reference
The organization of soil constituents into larger aggregates or compound particles.	The geometric or spatial arrangement of soil particles or voids.	Drees (1992)
The spatial arrangement and total organization of the soil system as expressed by the degree and type of aggregation and the nature and distribution of the pores and pore space.	The arrangement, size, shape, and frequency of the individual solid soil components within the soil as a whole and within features themselves.	Fitzpatrick (1993)
The gradation and arrangement of soil particles, porosity, and pore size distribution, bonding agents, and the specific interactions developed between particles through associated electrical forces.	The geometric arrangement of the constituents mineral particles, including the void space which can be observed visually or directly using optical and electron microscopic techniques.	Yong and Warkentin (1975)

in between the skeletal particles (Kubiens, 1938). The term “structural form” refers to the heterogeneous arrangement of solid and void space that exists in a soil at a given time. This term is used to describe arrangement of primary soil particles into hierarchical structural states (Kay, 1990). In contrast, an aggregate is a naturally occurring cluster of soil particles in which the forces holding the particles together are much stronger than the forces, between adjacent aggregates. Soil structure, however, is much more than a fabric or an aggregate. It is indeed hard to describe.

4.2 FORCES INVOLVED IN FLOCCULATION

There are several ionic forces involved in formation of floccules, domains, and aggregates (Fig. 4.2). Principal among these are inter and intramolecular forces, electrostatic, and gravitational forces.

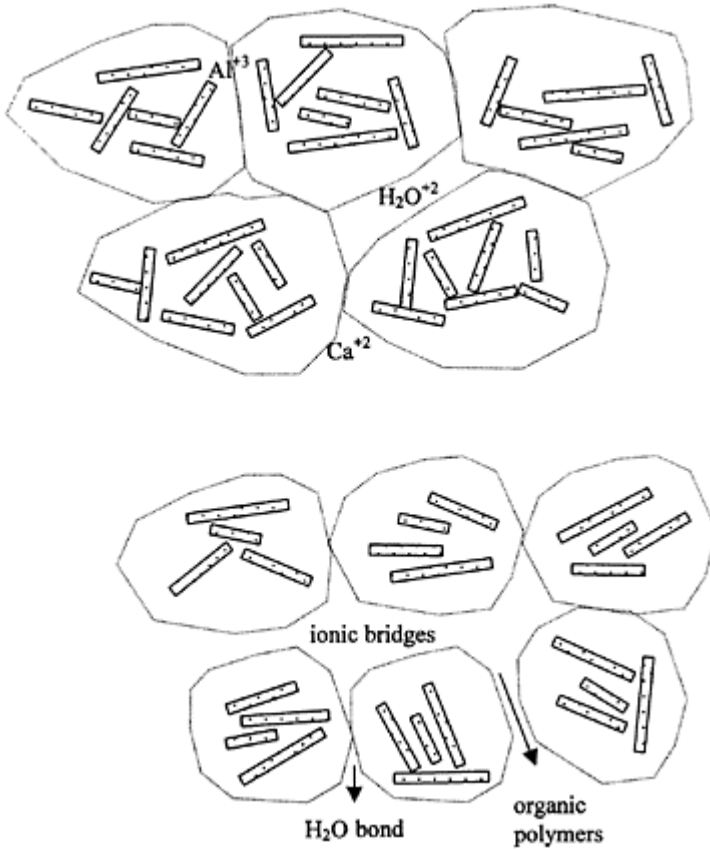


FIGURE 4.2 Floccules of clay particles cemented together lead to granulation. Aggregation is flocculation+cementation.

4.2.1 Intermolecular and Intramolecular Forces

Intermolecular attractions occur between one molecule with a neighboring molecule. The forces of attraction, which hold an individual molecule together (for example, the covalent bonds), are known as intramolecular attractions (see Chapter 3). All molecules experience intermolecular attractions, however, in some cases these attractions are weak. Even in a gas like H_2 , cooling slows the H_2 molecules down, and the attractions become large enough for the molecules to stick together to form a liquid and then a solid. For hydrogen, the attractions are so weak that the molecules have to be cooled to 21 K ($-252^\circ C$) before the attractions are enough to condense the hydrogen as a liquid. One type of intermolecular force is the van der Waals forces.

4.2.2 van der Waals Forces: Dispersion Forces

The weak forces that contribute to intermolecular bonding are known as van der Waals forces. These are weak attractive forces that hold nonpolar molecules together. The size of the attraction varies considerably with the size of the molecule and its shape. There are three types of van der Waals forces: intermolecular bonding, dispersion forces, and hydrogen bonding. The dispersion forces exist between nonpolar molecules and are also known as “London forces.” Hydrogen bonding is exactly the same as the dipole–dipole interaction that occurs between any molecule with a bond between a hydrogen atom and any of oxygen, fluorine, or nitrogen.

Water molecules in liquid water are attracted to each other by electrostatic or van der Waals forces. Even though the water molecule as a whole is electrically neutral, the distribution of charges in the molecule is not symmetrical and leads to a dipole moment—a microscopic separation of the positive and negative charge centers. This leads to a net attraction between such polar molecules, which finds expression in the cohesion of water molecules and contributes to viscosity and surface. The dipolar interaction between water molecules represents a large amount of internal energy and is a factor in water’s large specific heat (1 cal/g/°C or 1 cal/cm³/°C).

Nonpolar molecules also experience some van der Waals bonding, which can be attributed to their being polarizable. These molecules do not have permanent dipole moments, yet they can have instantaneous dipole moments, which change or oscillate with time. These fluctuations of molecular dipole moments lead to a net attraction between molecules, which allow nonpolar substances like carbon tetrachloride (CCl₄) to form liquids.

4.2.3 Electrostatic Forces

Electrostatic forces work in much the same way as magnetic forces, i.e., like forces repel and unlike forces attracts. Water molecules bond by an oxygen atom joining to one of the hydrogen atoms by means of a covalent bond where electrons are shared (refer to Fig. 3.16). The electrical force (F) is directly proportional to the product of the charges (q_1 and q_2) and inversely proportional to the square of the distance (r) between them.

$$F = \frac{1}{4\pi\epsilon_0} \frac{q_1q_2}{r^2}$$

where ϵ_0 is the permittivity constant and is equal to 8.854×10^{-12} Coulomb/(newton-m²).

The lack of electrostatic forces in everyday life reflects that matter consists of almost exactly equal numbers of positively charged protons and negatively charged electrons thoroughly intermingled with one another, mainly in the form of atoms. Electrons move around positively charged nuclei consisting of protons and neutrons. Electrons and protons have equal but opposite charges ($q=1.602 \times 10^{-19}$ Coulomb), and neutrons have zero charge. There is a perfect balance between the number of electrons and protons in ordinary matter, and the net charge is zero. Consequently, two separate objects near each other hardly exert any electrostatic force at all.

4.2.4 Gravitational Forces

Gravitational forces are always attractive, and 10^{36} times smaller than electrostatic repulsion between two protons. Gravitational forces involving massive objects can be strong enough to move Earth, and keep it in a nearly circular orbit around the Sun.

4.3 MECHANISMS OF AGGREGATION

The mechanism of aggregation involves exogenous driving forces and the endogenous interactive forces arising from the soil-water interaction. Consequently, the specific arrangement of soil particles as observed in the field is dictated by the nature of exogenous and endogenous forces involved. Advances in colloid chemistry have facilitated and improved our understanding of the mechanisms and processes of aggregation. The importance of clay and humus colloids in forming aggregates was recognized as early as 1874 by Schloesing. Dumount (1909) also pointed out the importance of amorphous colloidal material in aggregation. For details on earlier literature readers are referred to the review by Harris et al. (1966). Numerous theories have been proposed since the 1930s. For details on interparticle forces in relation to aggregation readers are referred to reviews by Murray and Quirk (1990), Oades (1990), Emerson and Greenland (1990), Tisdall (1996), and others.

4.3.1 Russell's Theory of Crumb Formation

Russell (1934) proposed that clay particles are bonded together into aggregates through ionic bonds (Fig. 4.3). The mechanism of crumb

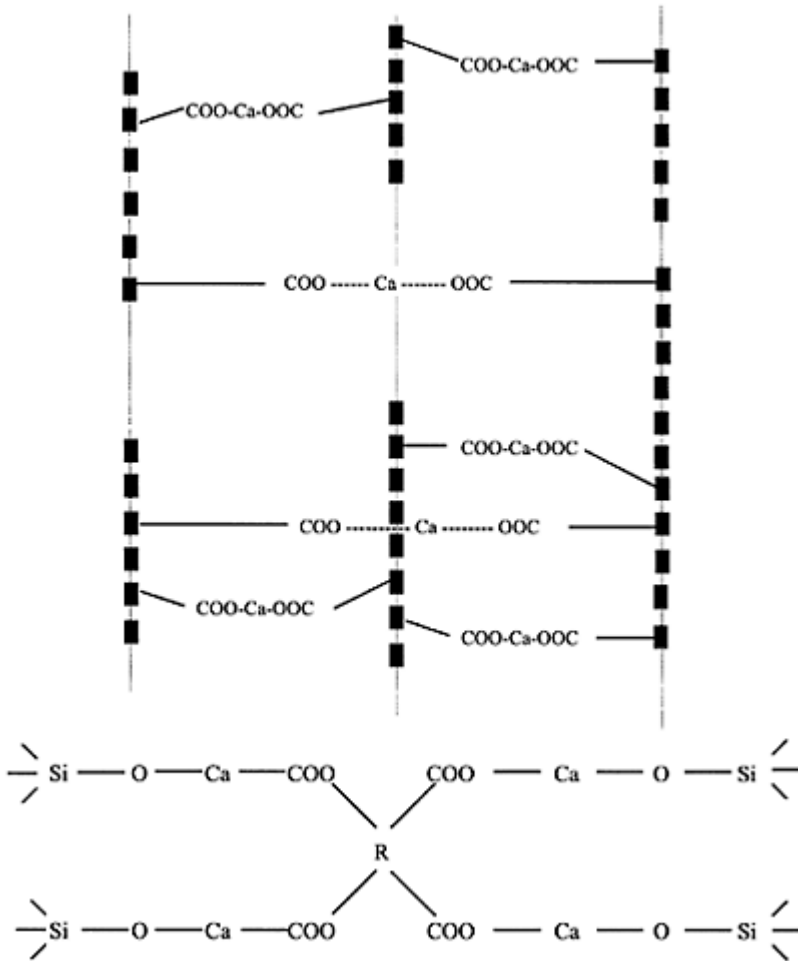


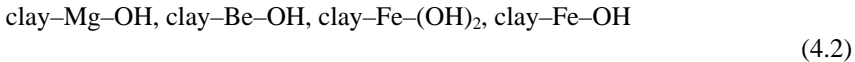
FIGURE 4.3 A hypothetical pattern for protopectin. (Redrawn from Peterson, 1947.)

formation according to this theory is as follows: (i) clay particles have a charge when hydrated, (ii) the charged particle is surrounded by an electric double layer of cations, (iii) polar water molecules are oriented along the lines of force radiating from each ion, and from each free charge of the clay particle, (iv) every clay particle is thus surrounded by an envelope of water, and (v) as the soil moisture content is reduced, the thickness of the envelope is reduced, and each ion shares its envelope with two clay particles thus holding the particles together. Russell observed that crumb formation according to this concept should meet the following requirements: (i) particles must have high cation exchange and large surface area, (ii) particles must be smaller than a certain size (1 μm)

because sand and silt fractions are not essential to crumb formation and make a crumb weaker, (iii) the liquid must have an appreciable dipole moment, and (iv) polyvalent cations must be present. Clay particles are adsorbed on sand and silt fractions, and the strength of bond between the clay and the sand increases with decreasing particle size of the clay. The process is reversible, because crumbs may disintegrate unless stabilized by appropriate cementing agents, because granulation is flocculation plus cementation.

4.3.2 The Calcium-Linkage Theory

Williams (1935) and Peterson (1947) proposed Ca-linkage as a mechanism in the formation of water-stable aggregates. The linkage was more effective in the presence of polyuronides, a component of soil organic matter, than without it. Negatively charged organic materials such as polysaccharides are adsorbed onto the surface of clay by Ca^{+2} or other polyvalent cations (Fe^{+3} , Al^{+3}). This model is schematically presented in Eq. (4.2) for different polyvalent cations, and Eq. (4.3) for Ca^{+2} , and schematically presented in Fig. 4.4.



4.3.3 Clay-Water Structure

Rosenquist (1959) proposed a concept of “clay-water structure.” Rosenquist suggested that adhesion between clay particles is based upon the difference in surface energy of the adsorbed water and the liquid pore water. Therefore, creation of interfacial tension between the two types of water may be the cause of cohesion observed in saturated clays. The concept of clay-water structure was also supported by the work of Lambe (1960), Michaels (1959), and Mitchell (1956).

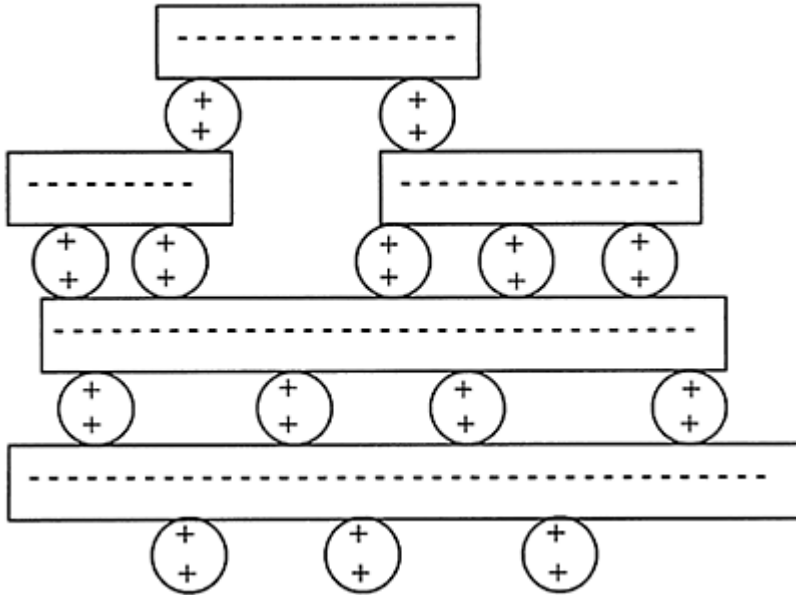


FIGURE 4.4 Plate-condensation of Ca-clay.

4.3.4 Edge-Surface Proximity Concept

Schofield and Samson (1954) and Trollope and Chan (1959) proposed a model based on the interparticle forces of attraction and repulsion. Their proposal of a card-house structure is based on the establishment of equilibrium between adjacent particles due to the edge-surface proximity establishing a link bond (Fig. 4.5). Flocculation occurs as a result of electrostatic attraction between the positive edges and negative faces of clay lattices. The link bond is established if the particles are sufficiently close to exceed the potential energy barrier. This model is essentially based on the forces of adhesion between the clay particles. This edge-to-face type of flocculation produces a much more stable system than flocculation caused by lowering of zeta potential due to addition of salt.

4.3.5 Emerson's Model

Emerson (1959) proposed that crumbs are formed by cementation of cardhouse or brush-heap type of floccules by positive edge-negative face attraction (Fig. 4.6). According to this model, both quartz and clay form the main components of an aggregate or crumb.

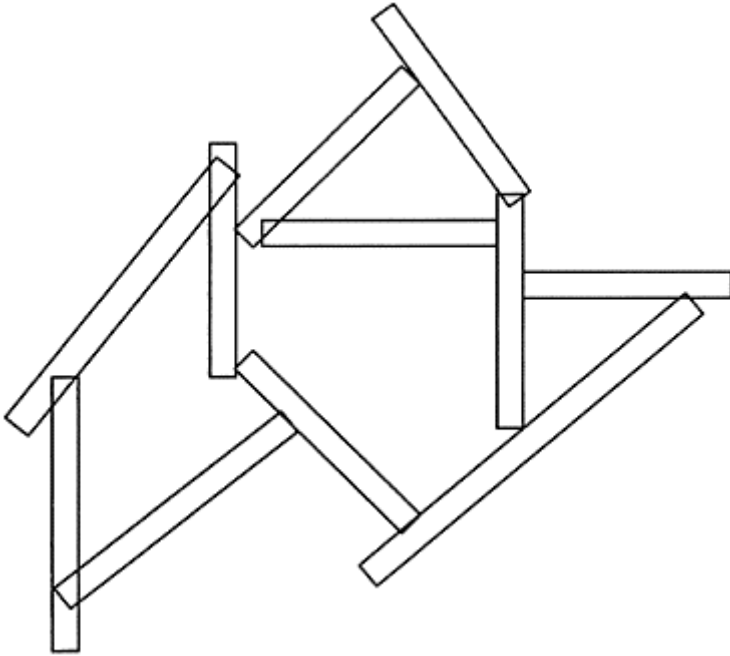


FIGURE 4.5 Card-house structure of floccules.

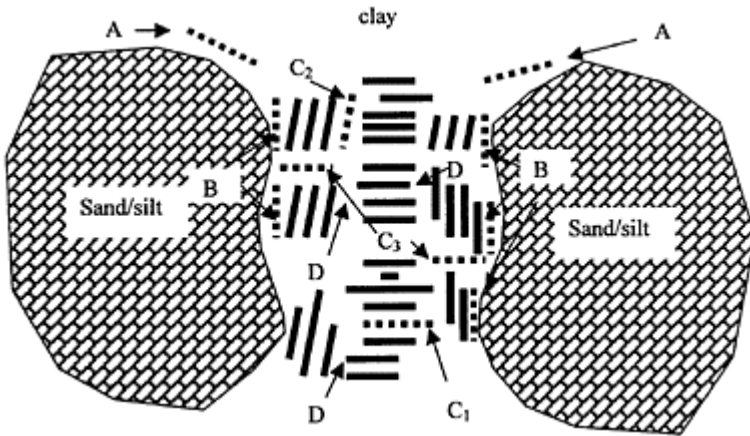


FIGURE 4.6 Schematic of the arrangements of quartz, clay domains, and organic matter in aggregate. Type of bond: A, quartz-organic matter-

quartz; B, quartz– organic matter-domain; C, domain-organic matter-domain (C₁, face-face; C₂, edge-face; C₃, edge-edge); D, domain edge-domain face. (Redrawn from Emerson, 1959.)

However, this structure dis-appears when soil is dried and 2:1 type clay minerals show an orientation with flat sides parallel. This crumb structure is generally stable when the exchange complex is dominated by Ca⁺² and other polyvalent cations. Emerson proposed four types of bonds prevalent in the crumb structure: (i) hydrogen bonding between the carboxyl group in organic matter and the clay, (ii) ionic bonding between the carboxyl group of organic matter and the clay, (iii) interaction of the electric double layers leading to the formation of domains, and (iv) bonding between the organic and inorganic colloids and between the colloids and the large soil particles. Emerson’s model is an extension of Russell’s model and incorporates the principles of the diffuse double layer. Clusters of clay crystals form domains as a result of orientation and electrostatic attraction to each other. These domains function as a single unit, and are bonded to the surface of the quartz grains and to each other to form aggregates. In addition, organic compounds increase the strength of the clay-quartz bond (Fig. 4.6). Electrostatic forces between the positive edges and negative faces of clay minerals, and presence of polyvalent cations also increase bond strength (Emerson and Dettman, 1960).

4.3.6 The Organic Bond Theory

Greenland (1965a; b) advanced Emerson’s model by showing the importance of soil organic matter in strengthening the bond between adjacent clay particles. Soil organic matter may hold particles together by ionic bonding in a manner similar to “string of beads.” For electrically neutral system, organic molecules may form a “coat of paint” around the outside of a number of particles binding them together into an aggregate.

4.3.7 Clay-Domain Theory

Williams et al. (1967) proposed that clay particles mostly exist in domains, up to about 5 µm in diameter, within which they are separated by “bonding pores” which maintain their identity. Clusters of domains are called microaggregates, with sizes in the order of 5–1,000 (µm, and microaggregates are clustered into aggregates, 1–5 mm in diameter (Fig. 4.7). The integrity of microaggregates and aggregates is dependent on cementation between domains or microaggregates by inorganic precipitates, or on organic materials acting as a lining spread over the surfaces of domains or microaggregates. Oriented clay films and microbial films may also bind microaggregates and aggregates.

4.3.8 Quasi Crystal Theory

Aylmore and Quirk (1971) extended Williams et al. (1967) domain model by introducing the concept of quasi crystals or packets. The latter involves parallel clay crystals (about 5 μm in diameter) which are clustered together

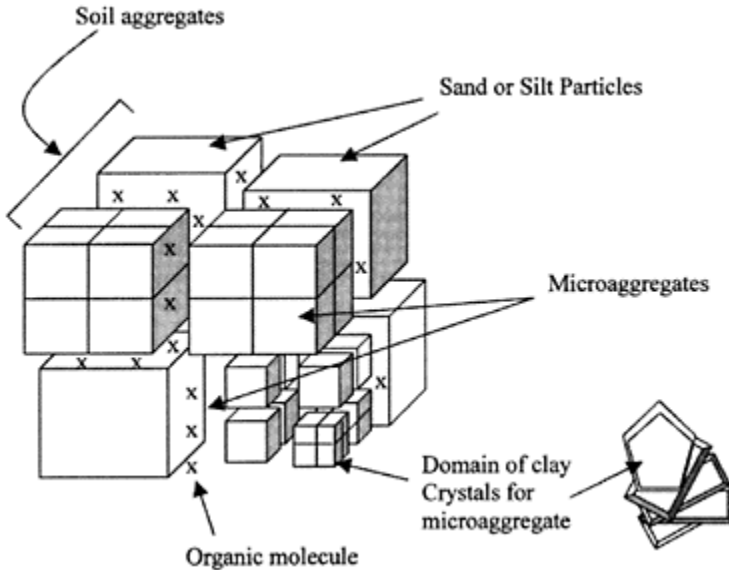


FIGURE 4.7 A hypothetical model of a soil aggregate. (Redrawn from Williams et al., 1967.)

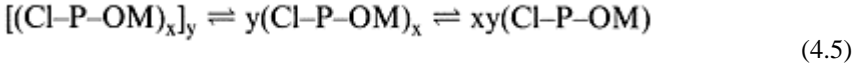
closely enough (0.01–1.3 μm apart) to form domains. Rather than using domains, Quirk and Aylmore proposed the term “quasi crystals” to describe the regions of parallel alignment of individual lamellae of aluminosilicates in swelling type clay minerals which exhibit the intracrystalline swelling (e.g., montmorillonite). In comparison, they used the term domain to describe the regions of parallel alignment of crystals with fixed lattice and which exhibit intercrystalline swelling only (e.g., illite). The quasi crystal model has been verified and supported by Oades and Waters (1991), who argued that clay particles are aggregated into quasi crystals or stable packets. Oades and Waters proposed three distinct size fractions: (i) binding of clay particles into stable packets <20 (μm), (ii) binding of clay packets into stable microaggregates 20–250 μm , and (iii) the binding of microaggregates into stable macroaggregates >250 μm .

4.3.9 Microaggregate Theory

Edwards and Bremner (1967) proposed that soil consists of microaggregates ($< 250 \mu\text{m}$) bound into macroaggregates ($>250 \mu\text{m}$), and bonds within microaggregates are stronger than those between microaggregates. Microaggregates are represented by the structure shown in Eq. (4.4).



where Cl is clay, P is polyvalent cation (Ca^{+2} , Al^{+3} , Fe^{+3}), and OM is organometallic complex including humified organic matter complexed with polyvalent metals. There may be more than one polyvalent metal bridge between clay (Cl) and OM in the Cl-P-OM units (Fig. 4.8). $(\text{Cl-P-OM})_x$ and $(\text{Cl-P-OM})_y$ represent compound particles of clay size ($< 2 \mu\text{m}$ in diameter) and x and y are finite whole numbers with limits dictated by the size of the primary clay particles. The bonds linking the Cl-P-OM clusters into the larger $(\text{Cl-P-OM})_x$ and $[(\text{Cl-P-OM})_x]_y$ units can be ruptured by chemical or mechanical treatments. Interparticle bonds are weakened by substitution of polyvalent cations by Na^+ (treatment with sodium hexametaphosphate) and by mechanical shaking (stirring) and ultrasound vibrations. However, reversal of the dispersion process can lead to the formation of stable microaggregates [(Eq. (4.5)].



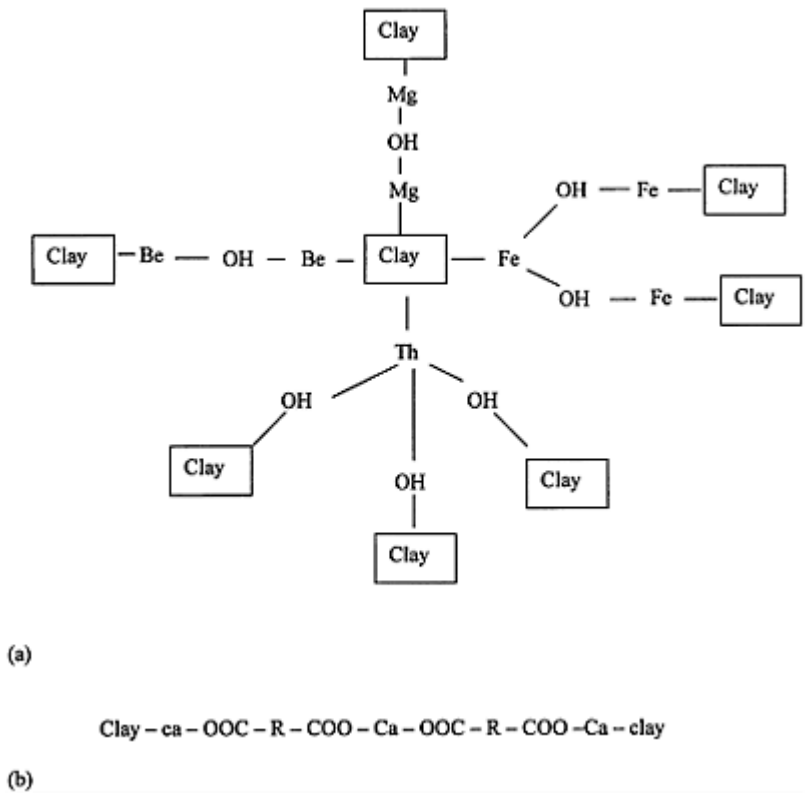


FIGURE 4.8 (a) Bridge between clay and polyvalent cations, (b) The calcium linkage between clay and organic polymers. (For details see Peterson, 1947.)

where D represent dispersion and A aggregation processes. This model has been verified by several researchers for Alfisols and Mollisols (Tisdall and Oades, 1982; Oades and Waters, 1991). Tisdall and Oades proposed that microaggregates themselves are built up in stages with different types of bonds at each stage (Tisdall, 1996; Table 4.1). Stages of aggregation are shown in Eq. (4.6)

$$\begin{aligned}
 &<0.2 \mu\text{m} \rightarrow 0.2\text{--}2 \mu\text{m} \rightarrow 2\text{--}20 \mu\text{m} \rightarrow 20\text{--}250 \mu\text{m} \\
 &\rightarrow >2000 \mu\text{m diameter}
 \end{aligned}
 \tag{4.6}$$

4.3.10 The Aggregate Hierarchy Model

Oades and Waters (1991) modified the stages proposed by Tisdall and Oades (1982) especially for soils whose aggregates are mainly stabilized by organic materials. The

modification was necessitated by the fact that it was not possible to distinguish steps of aggregation within aggregates less than 20 μm . They proposed that aggregates within the size range of 20–250 μm could be divided into aggregates 20–90 μm and 90–250 μm . Therefore, according to this model, the stages of aggregation or aggregation hierarchy are shown in Eq. (4.7):

$$<0.2 \mu\text{m} \rightarrow 20\text{--}90 \mu\text{m} \rightarrow 90\text{--}250 \mu\text{m} \rightarrow 250 \mu\text{m} \quad (4.7)$$

These aggregation hierarchies (Table 4.2) are developed over many years, and are, therefore, observed only in mature rather than young soils. Binding mechanisms for different size fractions are shown in Fig. 4.9.

4.3.11 The POM Nucleus Model

The hierarchy model presupposes different bonding mechanisms for different aggregate sizes, or spatial distribution and persistence of aggregating agents within the soil matrix. These bonding mechanisms include: (i) bonding of clay into quasi crystals or packets is governed by pedological processes through precipitates of sesquioxes as in Oxisols, and (ii) bonding of packets into microaggregates and aggregates is governed by various organic materials. The particulate organic materials (POM) form a nucleus or core around which clay packets and small microaggregates are bound into larger microaggregates (Elliot, 1996; Golchin et al., 1994) (Fig. 4.9). The POM is colonized by microbial population, and the microflora and its by-products have strong adhesive properties which bind the particles together (Lynch and Bragg, 1985). The plant fragments from

Table 4.2 Models of Aggregation and Major Stabilizing Agents

Soil type	Stabilizing agent	Stage of aggregation (μm)	Reference
Alfisol	Inorganic materials, organic polymers, electrostatic bonds, coagulation	<0.2	Tisdall and Oades, 1982
	Microbial and fungal debris	0.2–2→2–20	
	Plant and fungal debris	2–20→20–250	
	Roots and hyphae ^a	20–250→>2000	
	Ploysaccharides ^b	20–250→2000	
Alfisol, mollisol	Microbial debris, inorganic materials	<20	Oades and Waters,
	Plant debris	<20→20–90	
	Plant fragments	20–90→90–250	
	Roots and hyphae	20–250→>2000	
Oxisol	Oxides/sesquioxides	<20→>250	Oades and

Oxisol	Oxides/sesquioxides	<2→100–500	Waters, Robert and Chenu,
Vertisol	Organic matter	20–35→>250	Collis-George and Lal,
Andosols	Allophanes and amorphous aluminosilicates	0.001–0.01→01–1	Robert and Chenu, 1992

^aSoil with total organic carbon >2%.

^bSoil with total organic carbon <1 %.

Source: Adapted from Tisdall, 1996.

incorporation of crop residues, therefore, become the center of water stable aggregates (Buyanovsky et al., 1994; Angers and Chenu, 1997).

4.4 AGGREGATION AND STRUCTURAL FORMATION

Bradfield (1936) described that “granulation is flocculation plus.” He drew a sharp distinction between flocculation (see Chapter 3) and aggregation. The process of formation of soil aggregates or organomineral complexes, from primary particles and humic and other bonding substances, is called aggregation. It is the first step in the development of soil structure. The process of aggregation is closely linked with the behavior of the diffuse double layer and its response to ionic composition in the bulk solution (refer to Chapter 3).

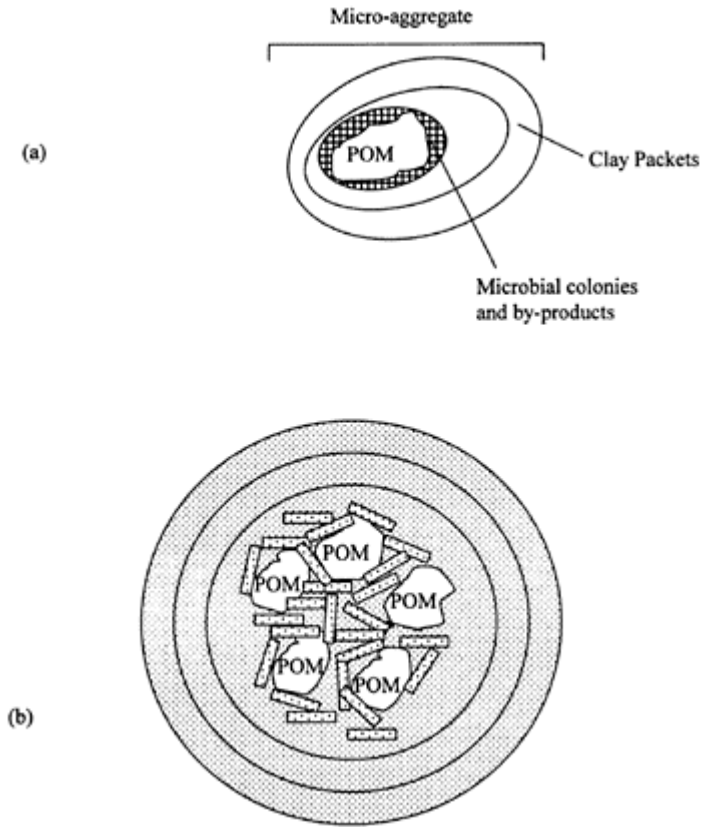


FIGURE 4.9 Microaggregates are formed around the particulate organic matter (POM) as a nucleus, (a) Microaggregate; (b) cluster of microaggregates forming a macroaggregate.

Aggregation is flocculation plus cementation with numerous forces, agents that stabilize and bind floccules [(Eq. (4.8))]:

$$\text{Aggregation} = \text{flocculation} + \text{cementation}$$

(4.8)

Most common cementing agents include soil organic matter, silicate clays, lime, and sesquioxide (FeO_3 , Al_2O_3 , Mn_2O_3) (Fig. 4.2). Humified organic matter, with its long polymer chains and electric charge balanced by polyvalent cations, is a very effective cementing agent. Fungal hyphae and microbial by-products also serve as cementing agents. In summary, there are four types of binding agents including: (i) oriented clay

films, (ii) microbial by-products, fungal filaments, and hyphae, (iii) inorganic precipitates such

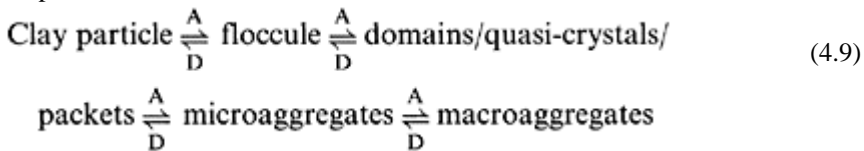
TABLE 4.3 Components of an Aggregate

Component	Size range
Clay	2 μm
Domain, quasi crystal, or Packets	2–5 μm
Microaggregate	5–500 μm
Aggregate	0.5–5 mm
Compound structure	>5 mm

as oxides of Fe and Al, and (iv) humic substances including organic polymers.

4.4.1 Bonding Agents Responsible for Aggregation and Structural Stability

Structural stability is the ability of a soil to retain its arrangement of solids and void space when external forces are applied. External forces may be natural or anthropogenic. The aggregate stability depends on the bonding agents involved in cementing the particles together. On the basis of the numerous models presented, components of an aggregate can be summed up as those shown in Table 4.3. The smallest component is domain or quasi crystal or packets. These are essentially floccules cemented together by different agents. The largest component is an aggregate that is < 5 mm. Anything larger than 5 mm may be a compound structure or a clod. Mechanisms of aggregation presented in the previous section can be summarized by Eq. (4.9) in which A denotes aggregation and D is dispersion.



There are different binding agents at each step going from clay particle to macroaggregates.

It has been argued that the reaction shown in Eq. (4.9) is as important as the photosynthesis reaction ($6\text{CO}_2 + 6\text{H}_2\text{O} \rightarrow \text{C}_6\text{H}_{12}\text{O}_6 + 3\text{O}_2$). Therefore, understanding the reaction in Eq. (4.9), and developing management strategies that push this reaction forward to the right-hand side are extremely important to crop production, and global food security.

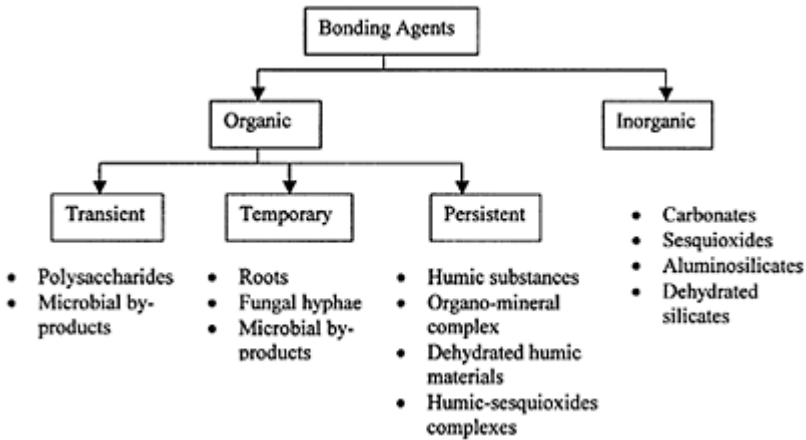


FIGURE 4.10 Different types of binding agents. (For details see Harris et al., 1966.)

The binding agents involved at each stage of aggregation can be grouped into three main categories described below and outlined in Fig. 4.10. For detail discussion on different binding agents, readers are referred to reviews by Harris et al. (1966) and Hamblin (1985).

Transient Binding Agents

These are organic materials that are decomposed very rapidly by micro-organisms. These materials include: (i) microbial polysaccharides produced when various organic materials are added to the soil, (ii) and some of the polysaccharides associated with roots and microbial biomass in the rhizosphere. These polysaccharides or glues are associated with large (>250 μm diameter) transiently stable aggregates, and are decomposed readily. Cellulose contributes to only a small fraction of aggregation but is more persistent. The transient polysaccharides (produced by bacteria, fungi, and plant roots) bind clay-sized particles into aggregates which are of the order of 10 μm diameter. Polysaccharides stabilize aggregates with diameter <50 μm .

Temporary Binding Agents

These agents are roots and mycorrhizal hyphae (Tisdall, 1991). Such binding agents are built up in the soil within a few weeks or months as the root system and associated hyphae grow. They persist for months or perhaps years, and are affected by management of the soil.

Roots. Roots supply decomposable organic residues to soil and support large microbial population in the rhizosphere. Roots of some plants,



FIGURE 4.11 Earthworm casts in a pasture enhance crumb structure and stable aggregates.

e.g., grasses, themselves act as binding agents. Residues released into the soil by roots are: (i) fine lateral roots, (ii) root hairs, (iii) cells from the root cap, (iv) dead cells, and (v) mucilages. The amount of organic carbon released by roots is proportional to the length of root. It can be 20–49 g of organic material per 100 g harvested root. The root system and associated hyphae of pasture plants, especially grasses, are extensive. The upper layer of the soil under pasture is probably all rhizosphere. Water stable aggregates are also formed due to localized drying around roots. Electron micrographs of a drying root show that particles of clay close to root tend to be oriented almost parallel to the axis of the root. Roots also provide food for soil animals, e.g., earthworms and the mesofauna. Population of earthworms in pastures may exceed $1.5 \times 10^6/\text{ha}$ (Fig. 4.11).

Hyphae. Hyphae are sticky and encrusted with fine particles of clay. Stabilization of aggregates by fungi in the field is limited to periods when readily decomposable material is available. Fungal hyphae are relatively large and usually bind microaggregates greater than 250 μm .

Saprophytic Fungi. This group of fungi includes dark colored fungi that tend to persist in soil.

Vesicular-Arbuscular (VA) Mycorrhizal Fungi. These are abundant in soils and are obligate symbionts. The VA mycorrhizal fungi tend to be most abundant in soils with low or unbalanced level of nutrients. Some plants are, however, mycorrhizal even in fertile soils. Mycorrhizal fungi bind particles into aggregates, and micro- into macroaggregates.

Other Temporary Binding Agents

Fungi constitute more than 50% of the microbial biomass in some soils and contribute more than bacteria to the organic matter in soil. Organic bonds also develop from degraded bacterial cells. In desert soils, filaments of blue-green algae are important. Algae and lichens form crust in desert soils.

Persistent Binding Agents

Persistent bonds include strongly sorbed polymers such as some polysaccharides and organic materials stabilized by association with metals. Degraded, aromatic humic materials associated with amorphous iron, aluminium, and aluminosilicates form the large organomineral fraction of soil that constitutes 52 to 98% of the total organic matter in soils. The persistent binding agents probably include complexes of clay-polyvalent metal—OM, C—P—OM, and (C—P—OM)_x both of which are <250 nm in diameter. Persistent binding agents are probably derived from the resistant fragments of roots, hyphae, bacterial cells, and colonies developed in the rhizosphere. The organic matter is in the center of the aggregate with particles of fine clay sorbed onto it, as opposed to the Emerson's concept of organic matter sorbed on the clay surface. Persistent bonding agents have not been defined chemically, just as the formula of humic acid cannot be defined. Some of these bonds resist ultrasonic vibrations.

The bonding forces in the formation of clay-organic complexes are summarized by Greenland (1965a). These forces are the same as those involved when atoms and molecules are in proximity. However, the situation is particularly complex when large organic molecules are involved. As is apparent from the discussion of various models of aggregation, the soil organic matter plays an important role in aggregation and structural stability of soils. It is not surprising, therefore, that numerous studies from around the world have demonstrated a high correlation coefficient between aggregation and soil organic matter content (Fig. 4.12). In contrast, there are also numerous studies indicating low or no correlation between soil organic matter content and aggregation. The lack of correlation, however, does not necessarily mean that soil organic matter content is not important to aggregation. The low or no correlation of aggregation with soil organic matter content may be due to several factors: (i) only part of the soil organic present is responsible for aggregation as is the case in soils of high organic matter content, (ii) there is a critical limit or threshold value of soil organic matter content above which it has no effect on aggregation, (iii) aggregation is affected by specific organic constituents rather than the bulk soil organic matter, (iv) there are other bonding mechanisms which are as good or more effective than soil organic matter, and (v) aggregation and aggregate stability are affected by other pedological or anthropological factors.

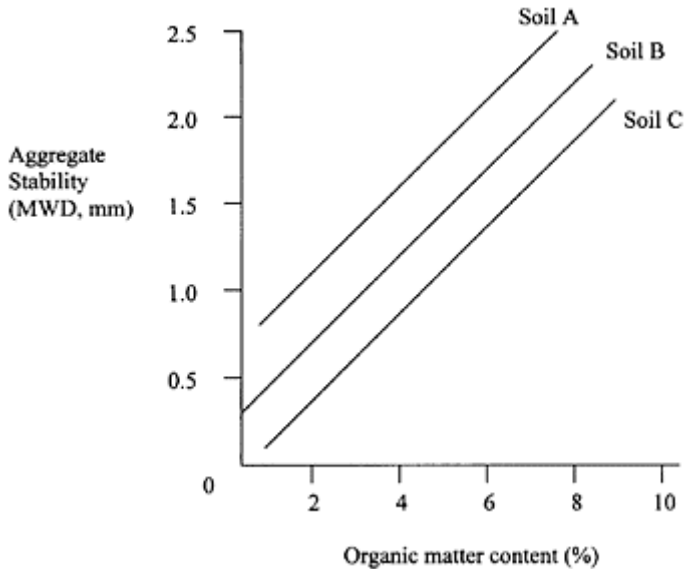


FIGURE 4.12 Schematic of the relationship between aggregate stability (mean weight diameter) and soil organic matter content for a group of soils.

In summary, there are different binding mechanisms for microaggregates and macroaggregates against rapid wetting and disruptive forces of cultivation and other natural or anthropogenic disturbances. Microaggregates are predominately stabilized by organo-mineral complexes. These bonds are relatively stable and not easily disrupted by changes in soil organic matter content brought about by land use and cultivation. In contrast, stability of macroaggregates depends on root hair and fungal hyphae. Therefore, the proportion of stable macroaggregates changes with change in soil organic matter content by land use and cultivation, and with changes in population of root hair and fungal hyphae. The stabilization of macroaggregates depends on management. It increases under fallow and pasture, and decreases with row cropping and plow-based tillage methods.

4.5 PROPERTIES OF AGGREGATES

An aggregate or ped thus formed is a distinct physical entity with quantifiable attributes, and exterior and interior properties. The exterior of an aggregate may be coated with: (i) clay film or “clay skins,” (ii) inorganic precipitates and sesquioxides, and (iii) organic matter. The exterior may have distinct shape (angular, subangular, prismatic, columnar, platy), size (coarse, medium, or fine) and strength or grade, and com-pactness. Similarly,

the interior of an aggregate may be compact or loose, anaerobic or aerobic, hygroscopic or hydrophobic, slow to dry when wet, or slow to wet when dry. Single aggregates are more dense compared to bulk soil (Horn, 1990; Kay, 1990). Bulk density generally increases with decrease in size of an aggregate (Becher, 1995). Two principal properties of an aggregate are strength and hydrophobicity.

4.5.1 Strength of Soil Aggregates

Strength refers to the ability of aggregates to withstand disruptive forces (e.g., vehicular traffic, raindrop impact, plowing, root pressure). The knowledge of magnitude and distribution of aggregate strength is key to understanding soil's response to tillage or traffic. Aggregated soils are stronger than nonaggregated or homogenized materials. Strength increases either by an increase in the total number of contact points between floccules and domains, or by increase in shear resistance per contact point (Hartge and Horn, 1984; Horn and Dexter, 1989; Horn et al, 1995). Factors affecting strength of soil aggregates are water content, texture, clay minerals, organic matter content and size of aggregates.

4.5.2 Hydrophobicity of Aggregates

Some coatings on aggregate surfaces impact their hydrophobic properties. Consequently, aggregates do not wet easily. Hydrophobic properties are attributed to some microbial by-products and other organic substances. In some soils, coverage of aggregates by such films is so extensive that water infiltration in soil is severely curtailed (see Chapter 14).

4.6 FACTORS AFFECTING AGGREGATION

There are numerous factors that affect aggregation (Hamblin, 1985; Kay, 1997) most of which can be grouped into two broad categories: endogenous and exogenous factors. The endogenous factors are those that are due to inherent soil properties. These factors include soil characteristics such as texture, clay mineralogy, nature of exchangeable cations, quantity, and quality of the humus fraction. The exogenous factors that affect soil structure include weather, biological processes, land use, and management.

The impact of seasonality, due to wetting and drying and freezing and thawing, on aggregation cannot be overemphasized (Bower et al., 1972). Biological processes, especially the activity and species diversity of soil fauna notably earthworms and termites, are extremely important to soil aggregation (Lal, 1987). Root growth is another important biological process affecting aggregation. Both of these exogenous and endogenous factors interact with one another, vary in both space and time, operate at different scales, and cannot be considered in isolation. Based on these and numerous interacting factors, there is a wide range of possible mechanisms and processes that lead to aggregation.

The literature is replete with analyses of factors affecting soil structure and strategies for its management (Bower et al., 1972; Kay, 1980; Hamblin, 1985; Carter and Stewart,

1996). Therefore, this section provides a brief outline of the salient features of the factors affecting aggregation under field conditions.

4.6.1 Drying and Wetting

Repeated cycles of drying and wetting play a major role in aggregation through shrinking and swelling that lead to formation of aggregates. Swelling or rewetting leads to reorientation of particles. Shrinking or drying leads to formation of cracks and increase in formation of link bonds through cementation. The mechanisms involved, especially the opposing forces, are not clearly understood. Non-uniform drying can lead to unequal strains throughout the soil mass. Consequently, large clods can break down into small aggregates by drying (Figs. 4.13a;b). Similar to rapid drying, rapid wetting also breaks large clods into aggregates because of the effect of entrapped air. That is why slow wetting, wetting by capillarity or wetting in vacuum is suggested for minimizing risks of soil slaking or rapid dispersion (Yoder, 1936; Henin, 1938). There is no slaking of aggregates if air in the soil is replaced by CO₂ (Emerson and Grundy, 1954; Robinson and Page, 1950). Other causes of slaking by rapid wetting include differential swelling (Panabokke and Quirk, 1957), and swelling of the oriented clay coatings or streaks (Brewer and Blackmore, 1956). However, the relative effectiveness of wetting and drying depends on the texture and cohesive properties of the soil (Grant and Dexter, 1989). In heavytextured soils, desiccation cracks lead to formation of ped faces (White, 1966; 1967). Rewetting of the shrunken soil causes swelling and development of shearing forces between the wet/dry boundary layer. Repeated shrinkage and swelling leads to formation of prismatic, blocky, parallelepiped, or platy peds in subsurface layers of heavy-textured soils.



(a)



(b)

FIGURE 4.13 (a) A freshly plowed field creates cloddy structure, (b) A weak structure creates surface seal that reduces infiltration rate. However, repeated wetting and drying cycles can improve aggregation.

4.6.2 Freezing and Thawing

Water expands on freezing, and its impact on aggregation depends on the size, distribution, and duration (or persistence) of ice crystals (Kay and Perfect, 1988). The in situ freezing of water in pores may lead to a fracturing of the soil. Local redistribution of water may also occur due to freezing



FIGURE 4.14 Repeated cycles of freezing and thawing also improve soil structure.

leading to accumulation of ice in large pores and shrinkage in adjacent areas. Large ice lenses are formed when large quantities of water move from the unfrozen zone up into the frozen zone in response to freeze-induced gradients in soil-water potential. Ice lenses may cause formation of a laminar structure in a silt loam but a distinctly reticular or polygonal structure in a clay loam soil (Ceratzki, 1956; Kay et al., 1985). The most important effect on aggregation is of the cyclic freezing and thawing (Pawluk, 1988) (Fig. 4.14). Fabric changes occur in plastic clays by freezing and thawing (Czurda et al., 1995). Despite numerous observations on the positive effects, Slater and Hopp (1949) and others have reported negative effects of freezing on structural attributes. An important factor determining the effect is the degree of soil wetness at the time of freezing (Logsdail and Webber, 1959), and number of freeze-thaw cycles. There appears to be a maximum in the positive effects of freeze-thaw cycles.

4.6.3 Biotic Factors

Soil biota plays an important role in aggregation and soil structure development (Fig. 4.15). In addition to the significant effects of plant roots, soil fauna drastically alters soil structure (Lal and Akinremi, 1983; Lal et al., 1980; Lee, 1985; Lal, 1991; Lavelle and Pashanasi, 1989; Lee and Foster, 1991; Schrader et al., 1995). The role of root hairs, fungal hyphae, and other mineral by-products of soil biota have been discussed in the previous section. Enhancing microbial activity in soil is an important strategy of improving soil structure. Products of microbial decomposition facilitate clay-organic complex formation.

4.6.4 Soil Tillage

Shearing, compressive, and tensile stresses during seedbed preparation drastically alter porosity and pore size distribution due to change in soil



FIGURE 4.15 Termite activity is more predominant in tropical than temperate region soils, and their activity creates aggregates and channels.



FIGURE 4.16 A wheel rut causes soil compaction.

volume (Spoor, 1988). Wheel traffic has a significant effect on soil structure (Fig. 4.16; Hakansson et al., 1988). Conservation tillage (Fig. 4.17), and use of crop residue mulch (Fig. 4.18), is an important strategy to maintain a favorable structure of some soils.



FIGURE 4.17 No-till farming with residue mulch enhances activity of soil microfauna (e.g., earthworms, termites) and improves soil structure.



FIGURE 4.18 Crop residue mulch, in situ or brought in, also improves soils structure by eliminating the raindrop impact and enhancing activity of soil macrofauna. (The pen points to earthworm casts beneath the mulch layer.)

4.6.5 Soil Amendments

Addition of organic matter (e.g., compost, manure, sludge) has beneficial effects on soil structure through formation of clay-organic complexes (Greenland, 1965a; b; Glass, 1995). Similarly, application of gypsum (CaSO_4) leads to improved aggregation of dispersed alkaline soils ($2\text{Na}^+ + 1\text{-clay} + \text{CaSO}_4 \rightarrow \text{Ca-clay} + \text{Na}_2\text{SO}_4$) (Gupta and Abrol, 1990). There are also synthetic organic polymers or soil conditioners or soil stabilizers (Levy, 1996). In fact, interest in organic polymers as soil conditioners dates back to the 1950s when the Monsanto company developed Krilium, a trade name comprising several polymers such as vinyl acetate, malic acid, and hydrolyzed polyacrylonitrile (Chepil, 1954; De Boodt and De Leenheer, 1958; Emerson et al., 1978). Polymers are small repeating units or monomers coupled together to form extended chains. Their chain length in solution ranges between a few thousand and $3 \times 10^5 \mu\text{m}$ with an average diameter

of 0.5–1.0 μm . Commonly used polymers are polysaccharides (PSD) and polyacrylamide (PAM). Clay–polymer complexes lead to formation of stable aggregates (De Boodt, 1972; Gabriels et al., 1973; SSSA, 1975). The cost-effectiveness and the persistence of the effect need to be carefully assessed under soil/site specific situations.

4.7 ASSESSMENT OF AGGREGATION AND SOIL STRUCTURE

Soil structure is a dynamic property with numerous aspects, and is difficult to characterize (Coughlan et al., 1991). Methods of aggregation assessment outlined in Fig. 4.19 show two principal techniques: field and laboratory.

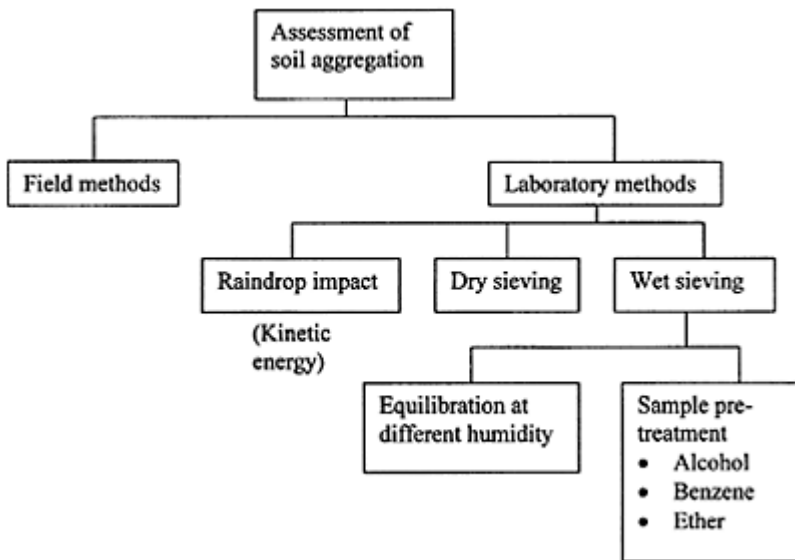


FIGURE 4.19 Methods of assessment of aggregation.

Field methods are primarily used by pedologists in routine soil surveys (Soil Survey Division Staff, 1993).

4.7.1 Pedological Methods

Soils are classified into structureless and structured soils. Structureless soils may either be single-grained such as sand or massive such as large clods without distinctive peds. No peds or units are observed in structureless soils. Structured soils may have simple or compound structure. Simple structure comprises distinct aggregates, which are an entity unto themselves without components or smaller units separated by persistent planes of weakness. Zakhrov (1931) described soil structure based upon the size, shape, and visual

appearance of the surface of soil aggregates, fragments, or clods. Soil type refers to shape, size to class, and grade to durability of peds (Table 4.4). The Soil Survey Division of the Soil Conservation Service (now called Natural Resource Conservation Service) of the USDA revised the Zakhrov system. The revised version of the field/pedological method is shown in Fig. 4.20 and Table 4.4. Morphologic features of struct-ural units are also classified based on soil fabric involving petrographic studies (Kubiens, 1938; Brewer and Sleeman, 1960; Brewer, 1964; Ringrose-Voase, 1991).

Table 4.4 Zakhrov System of Classification of Soil Structure

Type Criteria	Form	Types	Size (mm)
1 Structure develops uniformly along three mutually perpendicular axes (polyhedral or round)	a. Faces and edges not well defined	1. Lumpy	50–100
	b. Faces and edges are well defined	2. Crumbly	5–50
2 Structure develops more toward the vertical axis (prismatic)	a. Rounded apexes b. Apexes bounded by plane facets	1. Nuciform	30–50
		2. Grainy	10–50
3 Structure develops along the horizontal axis (platy)	a. Well-developed horizontal cleavage	1. Columnar	1–5
	b. Cleavage planes bent horizontally	1. Prismatic	<1
	c. Top and bottom bound by round surfaces	2. Leafy	>3
		1. Concoidal	1–3
		2. Flaky	3–10
		1. Lenslike	<3
		2. Lenticular	

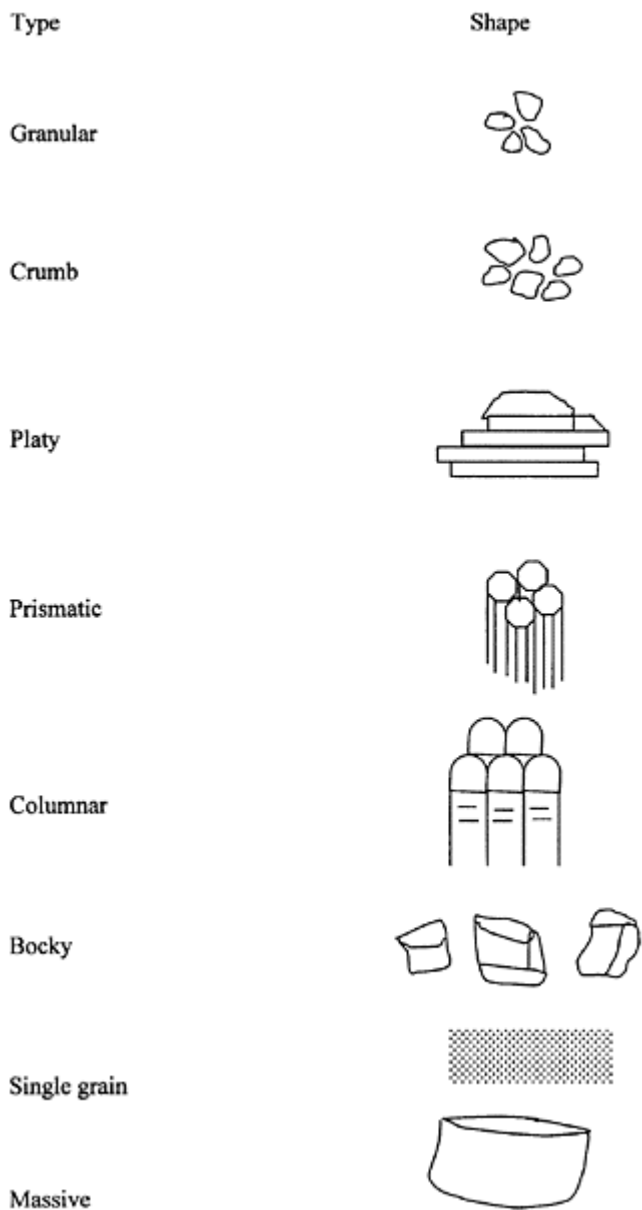


FIGURE 4.20 Classification of soil structure according to shape.

4.7.2 Laboratory Methods

Several books have been written in describing laboratory techniques of soil structure evaluation (Lorinov, 1982; Revut and Rode, 1981; Burke et al., 1986; Hartge and Stewart, 1995). Laboratory methods of aggregate analyses can be broadly grouped into three categories: (i) ease of dispersion, (ii) assessment of aggregation and aggregate size distribution, and (iii) evaluation of aggregate strength. Different methods are outlined in Table 4.5.

Dispersion

A known quantity of air dry soil is poured into a beaker containing deionized or distilled water. Quick wetting of aggregates leads to aggregate breakdown. Emerson (1967) developed a classification of soil aggregates based on their coherence in distilled water as judged by slaking and dispersion. Turbidity of water is measured as an index of ease of dispersion or slaking of aggregates (Emerson, 1954; 1964; 1967). Several indices have been developed to classify soils on the basis of their dispersion characteristics (Janse and Koenigs, 1963).

Aggregation and Aggregate Size Distribution

Resistance of soil solids to the mechanical abrasion arising from the movement of the solids relative to the surrounding medium (water or air) has long been used to measure stability of aggregates. Wet sieving analysis has long been used in evaluating the water stability of aggregates (Tiulin, 1928; Yoder, 1936). Wet sieving may be done with and without pretreatment of the samples to evaluate the relative importance of different binding agents (Henin et al., 1959; De Leenheer and De Boodt, 1959; De Boodt and De Leenheer, 1958). While wet sieving is done to simulate erosion by water and stability to quick wetting, dry sieving is done to simulate aggregate resistance to wind erosion. The techniques for aggregate analysis are described by Kemper and Rosenau (1986). Wet sieving techniques are discussed by Angers and Mehuys (1993) and dry sieving by White (1993). In highly aggregated soils, ultrasonic vibrations have been used to determine aggregate stability under wet conditions (North, 1979). The dispersive energy per unit mass of soil is related to aggregate stability.

Aggregate Strength

Aggregate strength may be determined by the raindrop technique (McCalla, 1944; Bruce-Okine and Lal, 1975) by evaluating the kinetic energy required to disrupt an aggregate. Dry soil aggregate strength may be evaluated by a procedure that evaluates crushing strength (Skidmore and Powers, 1982; Perfect and Kay, 1994). A soil energy-crushing meter has been developed (Boyd et al., 1983).

Table 4.5 Shapes and Size Classes of Soil Structure

Size class	Platy (mm)	Shape of structure				
		Prismatic (mm)	Columnar (mm)	Angular blocky (mm)	Subangular blocky (mm)	Granular (mm)
Very fine or very thin	<1	<10	<10	<5	<5	<1
Fine or thin	1–2	10–20	10–20	5–10	5–10	1–2
Medium	2–5	20–50	20–50	10–20	10–20	2–5
Coarse or thick	5–10	50–100	50–100	20–50	20–50	5–10
Very coarse or very thick ^a	>10	>100	>100	>50	>50	>10

^aIn describing plates, *thin* is used instead of *fine* and *thick* is used instead of *coarse*.

Source: Soil Survey Staff, 1951; 1993.

4.7.3 Expression of Results of Aggregate Analysis

Numerous methods are used to express the results of structural analysis (Table 4.6), and there are different methods to express results of aggregate analysis (Table 4.7). Commonly used methods to express results include percent water stable aggregation

(%WSA) and mean weight diameter (MWD) of aggregates (Van Bavel, 1949; Youker and McGuinness, 1956). It is important that the MWD is corrected for the primary particles of the same size to avoid over-estimation of the MWD. The correction in MWD for sand is done as per Eq. (4.10).

% Stable aggregates on each sieve

$$= \frac{(\text{weight retained}) - (\text{weight of the sand fraction})}{(\text{total sample weight}) - (\text{weight of sand})} \times 100 \quad (4.10)$$

Results of aggregate analysis are also expressed as geometric mean diameter or GMD (Table 4.7). In general, GMD is lower numerically than MWD.

4.7.4 Indices of Soil Structure

There are also several other indices of soil structure based on soil properties other than aggregation. Important among these are those based on porosity, soil strength, plant available water capacity, and water transmission properties. These indices are outlined in Table 4.8 but discussed in detail in appropriate chapters.

Rather than doing the direct evaluation for total aggregation, and their size distribution and strength, there are numerous indirect indices of soil structure assessment. These indices are based on other soil properties related to soil structure, and have been described by Bryan (1968). Some of these indices include the following:

Dispersion Ratio (Middleton, 1930)

This index is a measure of the clay fraction in dispersed rather than aggregated condition. The dispersion ratio (*DR*) index is given by Eq. (4.11).

$$DR = \frac{a}{b} \times 100 \quad (4.11)$$

where *a* is percent (silt+clay) when 50 g of oven dry equivalent sample is mixed end over end without dispersion agent in one liter of distilled water

TABLE 4.6 Methods of Determining Structural Stability

Method	: Formula/technique	Reference
1. Slump test	$\frac{Z - Y}{Z - X} \times 100$	Z=Initial volume of soil column Y=Final volume of soil column X=Absolute volume of solids William and Cook (1961)
2. Turbidity/slaking test	Turbidity classes	Panabokke and Quirk (1957) Quirk and Panabokke

		(1962) Emerson (1967) Janse and Koenigs (1963) Molope et al. (1985) Pajasok and Kay (1990)
3. Thorburn subsoiling test	Amount of cold dispersed in water	Thorburn (cited by Emerson, 1967)
4. Stability against water or wind	Wet and dry sieving	Yoder (1936)
5. Aggregate strength	(i) Kinetic energy (ii) Crushing strength (iii) Rupture energy	Bruce-Okine and Lal (1975) Skidmore et al. (1982) Perfect and Kay (1994)

TABLE 4.7 Some Commonly Used Indices to Express Results of Aggregate Analysis by Wet or Dry Sieving

Mean weight diameter	$\text{MWD (mm)} = \sum_{i=1}^n x_i w_i,$ <p>where x_i is mean diameter of each size fraction (mm) and w_i is proportion of the total sample weight occurring in the corresponding size fraction, and n is the number of size fractions.</p>
Geometric mean diameter	$\text{GMD} = \exp \left[\frac{\sum_{i=1}^n w_i \log x_i}{\sum_{i=1}^n w_i} \right],$ <p>where x_i is mean diameter of each size fraction (mm), w_i is weight of aggregate in a size class with an average diameter x_i (g), and $\sum_{i=1}^n w_i$ is the total weight of the sample.</p>
Distribution percent by	$DPW = S_c/S_o \times 100$, where S_c is oven dry weight of soil remaining on weight.
Percent silt plus clay aggregated	$PSC = \frac{W_{ag} - W_p}{W_s} \times 100$, where W_{ag} weight of aggregated soil, W_p is weight of soil particles retained on 0.02 mm sieve, and W_s is weight of original oven dry soil.
Percent clay aggregated	$PC = \frac{W_d - W_{nd}}{W_d} \times 100$, % clay with dispersion, W_{nd} is % clay without dispersion
Summation curve Log normal statistical	Cumulative % is plotted as a function of the aggregate size. The DPW is plotted on y-axis on a linear scale and the aggregate size on x axis in the log