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Advances in Characterization of Soil Structure

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ABSTRACT

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Soil structure is defined as "the spatial heterogeneity of the different components or properties of soil". Aspects of soil structure which are important for plant development, soil water balance and soil workability are reviewed briefly. The different types of soil structure which occur on different size scales are placed in a hierarchical order. Different mechanisms give rise to the different hierarchical orders. Similarly, different physical/chemical/biological processes are involved in the stabilization of the different hierarchical orders.

A number of methods for measuring soil structure are described. Preference is given to methods involving direct observation of structural features by scanning electron microscopy and by optical scanning of impregnated sections and fracture surfaces. These need to be supported by assessments of the stabilities of compound particles in water and of the mechanical strengths of compound particles as a function of water content.

"Good" soil structure is described as one where all the hierarchical orders are well-developed and stable. The greatest lack of knowledge appears to be in the 2-100 μm size range which is too large to have been studied by colloid chemists and too small to be visible to the naked eye. It is suggested that more observations of soil structure should be made in this size range, as it may hold many important clues on how to manage soil structure in the field.

INTRODUCTION

Soil structure may be defined as "the spatial heterogeneity of the different components or properties of soil". This definition accommodates the many different aspects of soil structure which are manifest at many different size scales in soil. Therefore, the arrangement of colloidal clay particles in a floccule; the arrangement of clods on the surface of a tilled layer; an array of earthworm tunnels; and the variability of soil strength from one point to another are all aspects of soil structure. In other words, spatial heterogeneity = spatial variability = structure.

The range of size scales involved in soil is enormous. In his everyday tasks, a soil scientist may consider fine clay particles with a typical dimension of 10^{-7}

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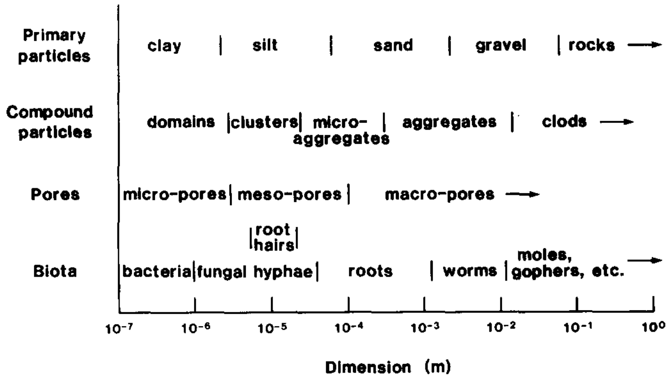

Fig. 1. Approximate dimensions or typical diameters of common soil structural features. Some primary particles and pores between clay plates may be two orders of magnitude smaller than shown.

m and variability of soil properties across one or more fields over distances of, say, 10^3 m. This gives a factor of 10^{10}. One need not stop there: one could consider atomic dimensions and inter-continental variability and get a factor of 10^{17}. The latter would be rather extreme, but it serves to illustrate the point that it is unlikely that any single experimental method or soil physical process will be applicable over the whole range of size scales. Therefore, we need different methods to study the different structural features and physical processes which occur at the different size scales.

Several different categories of soil features have overlapping size ranges. Thus, primary particles (i.e. single mineral crystals or grains) may be of the same size as compound particles, which are composed of many smaller primary particles. Some sizes and names of size categories are shown in Fig. 1. Also included in the figure for comparison purposes are some approximate dimensions of soil biological organisms or biota.

Soil structure has significant effects on plant development, soil water balance and soil workability. The following sections include brief reviews of these topics as well as some concepts of soil structure and some ideas about structure development, the stability of soil structure and methods for measurement of soil structure.

SOIL STRUCTURE AND PLANT DEVELOPMENT

Water supply to roots

Plant roots need supplies of both water and oxygen and both of these have to be supplied by or through the soil. The maximum amount of water that soil can store for a reasonable length of time (i.e. several days) is known as the "field capacity". Unfortunately, this value is not unique, especially for heavy clays which do not reach an equilibrium water content within a reasonable drainage time. For many practical purposes, however, field capacity corresponds to a water matric potential of around -10 kPa, although soil scientists in various countries use values in the range -6 to -33 kPa. At a potential of -10 kPa, water fills all pores up to 30 μm diameter (equivalent cylindrical diameter), except "blocked" pores, which contain entrapped air. At this potential, all pores >30 μm (equivalent cylindrical diameter) are filled with air.

Most agricultural crops can extract water from soil until the soil water potential is around -1.5 MPa, at which point the plants wilt. At this potential, and ignoring any osmotic components, all pores up to 0.2 μm diameter are full of water. Therefore, the maximum possible storage of water for plant use occurs when the volume of pores between 30 and 0.2 μm diameter is maximum.

Water supply to roots requires not only storage capacity but also the ability of the soil to transmit water to the root surfaces in response to potential gradients. Minimum hydraulic conductivity of the bulk soil must be around 10^{-4}–10^{-5} mm day^{-1} if water supply is not to restrict plant development (Newman, 1969; Taylor and Klepper, 1975; Reicosky and Ritchie, 1976; Hasegawa and Sato, 1985).

In Urrbrae loam at the Waite Institute, for example, hydraulic conductivities of 10^{-4} and 10^{-5} mm day^{-1} normally occur at approximate matric water potentials of -1.0 and -2.6 MPa, in the A-Horizon (17% clay) and at -0.4 and -1.8 MPa in the B-Horizon (60% clay). Therefore, except perhaps under high-transpiration conditions, the ability of this soil to transmit water to the roots is not a limitation even at water contents down to the wilting point.

Aeration

Maximum water storage is not the only criterion, however, because of the aeration requirement. Plant roots need oxygen in quite large amounts, and oxygen moves extremely slowly through water. Therefore, continuous, air-filled pores are required in the soil down to the optimum depth of rooting of the plants. Typically, the requirement for plant development is for at least 10% of the soil volume to comprise gas-filled pores at field capacity, and for at least 10% of the gas in these pores to be oxygen. Of course, the requirement is for a certain oxygen supply to the plant roots and not for a given average level in the soil. For this reason, oxygen supply to roots depends on many complex factors including pore continuity, tortuosity, sizes and spacings of air-filled pores.

Oxygen consumption by soil microorganisms can be similar in magnitude to the requirement of the crop. Microbial oxygen consumption can result in anaerobic conditions at depth or in the centres of dense clods or peds. Anaerobic conditions can inhibit root development except for certain hydrophytes, such as rice.

So, at least in simple terms, the requirement is for 10% of the soil volume to be in pores larger than 30 μm for aeration, and a maximum volume of pores between 30 and 0.2 μm for water storage.

In soils which do not shrink or swell, water drainage or extraction by roots results in an increase in soil air-filled porosity because the total pore volume is constant. In soils which shrink and swell, this is not the case and air-filled porosity of the peds between any shrinkage cracks can remain essentially constant or increase by only a small amount as water is removed. As a shrinking soil dries and cracks, the cracks may provide well-aerated pore space, but the large peds between the cracks may still be anaerobic and unsuitable for root growth. For this reason, only a small proportion of the total soil volume may be exploitable by roots in swelling/shrinking soils.

For recent reviews of the important subject of aeration, the reader is recommended to examine Glinski and Stepniewski (1985) and Edling (1986).

Seedbed conditions

Seeds need to be in a certain soil structural environment for optimum germination and early growth. Several researchers have shown that the optimum seedbed is composed of aggregates with a range of sizes between 1 and 5 mm diameter in the vicinity of the seed (e.g. Russell, 1973). For crops with very small seeds, smaller aggregates than this may be necessary to prevent the seeds from falling down through the inter-aggregate pores. However, it is important that there is always $<15\%$ of fine material (<250 μm) which can block the larger pores. The optimum seedbed provides adequate soil–seed contact for water supply for swelling and germination, and also adequate aeration. Sometimes, light rolling with rollers or press-wheels is necessary to get adequate soil–seed contact. However, the application of excessive pressure can lead to compaction of the seedbed, which can delay germination (Hadas, 1985).

This optimum environment for germination and early growth, of course, needs to exist only in the vicinity of the seeds. It is not necessary for the whole field to be in this condition. The seedbed may need to have larger aggregates nearer to the surface, especially for prevention of water and wind erosion. Uniform seeding depth and seed environment are required for uniform crop development.

Smaller aggregates and/or pores may exist below the depth of seeding. There is an optimum soil bulk density for the layer between the depth of seeding and the depth of ploughing (e.g. 50–250 mm depth). Håkansson (1987) has shown that this density can be simply expressed in terms of the density obtained in a standard compaction treatment.

The principal physical properties of seedbeds have been identified in experiments on beds of sieved aggregates (Håkansson and von Polgár, 1976, 1977, 1979, 1984). Measurements of the structures of real seedbeds in the field have been made by Kritz (1983) who used a sieving technique, and by Dexter et al. (1983) and Adem et al. (1984) who made statistical analyses of the distribution of aggregates and pores as observed on sections cut through blocks of impregnated seedbeds.

Mechanical constraints

Plants require only enough roots for the uptake of the water and mineral nutrients needed for their development. In most soils and climates this requires a high density of roots near the surface where the concentrations of most of the mineral nutrients are greatest, and at least one root per plant penetrating deep into the sub-soil so that the plant can extract the water stored there and possibly also the nitrate which may have been leached there from above.

For many crops in good soils, normal root density profiles are approximately exponential with the highest density close to the surface. Soil layers which present constraints to root growth may distort the root density profile from the normal. Root densities in the top soil can be very high with values of $L_v = 10\ cm\ cm^{-3}$ for cereals which corresponds to a mean spacing between root axes of only about 3 mm (Barley, 1970). Rarely, however, do roots occupy more than 2% of the total soil volume. Root densities in the subsoil (1 m depth) may be only $L_v = 1\ cm\ cm^{-3}$ or less. For many crops, root growth ceases completely at some depth between 1 and 2 m.

In order to develop such a root system, a plant must either exploit existing soil pores of suitable size or must make new pores (root channels) by overcoming the soil strength and pushing the root tips forward. Root tips are usually unable to enter rigid pores which are smaller than about two thirds of the diameter which they have when they are grown under zero stress conditions (e.g. in solution culture). However, if the soil is deformable, they may be able to enlarge existing pores while still taking some advantage from them. Root hairs (which are important in nutrient uptake) are thought to be unable to form their own pores, and therefore require pre-existing pores equal to or larger than themselves (typically 10 μm diameter) (Champion and Barley, 1969).

Root tips osmoregulate against the mechanical stress which they have to exert to penetrate the soil. They are thought to do this linearly and to be able to exert a maximum pressure on the soil of around 1 MPa (Dexter, 1987a). The whole subject of root growth in relation to soil strength has recently been reviewed by Greacen (1987). The expenditure of energy by the plant in overcoming soil strength is negligible, but the effects of soil strength on root elongation rate are significant.

Soil strength, as perceived by a root of a certain size, is a consequence of the soil structure on smaller size scales. This is because strength is strongly influ-

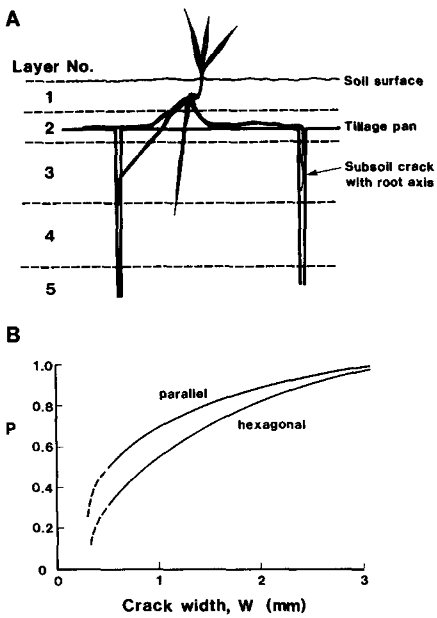

Fig. 2. Effect of vertical cracks on root growth. A shows a vertical section of a soil profile where a tilled layer overlies a hard, compacted layer. Roots may be deflected horizontally until they reach the cracks (after Jakobsen and Dexter, 1987). B shows the proportions of deflected roots which are able to enter the vertical cracks as functions of crack width and cracking pattern (after Dexter, 1986b).

enced by the packing of the soil particles and by the "cementing" between the particles which can be produced by smaller particles (e.g. fine clay) and by various inorganic and organic compounds.

Structure on size scales larger than a root can also influence the behaviour of a root. For example a compacted layer at the base of the seedbed or tilled layer can inhibit or prevent root penetration. In this case, root axes may be deflected to grow horizontally along the top of the compacted layer until the root may encounter a vertical pore of sufficient size to accommodate it. In this way, roots may be able to by-pass compacted soil to reach weaker soil layers, perhaps containing supplies of available water, which may exist beneath it. With the idea of making artificial cracks Elkins and Hendrick (1982, 1983) have developed the concept of "slit tillage". In this, slits of about 3 mm wide are made through a hard-pan and provide pathways for deep root penetration. Experimental results and simple mathematical models for such cases have been presented by Dexter (1986a,b,c) and some results are shown in Fig. 2.

When roots are confined to cracks between soil peds, water extraction occurs only from the ped faces (Hasegawa et al., 1985). Effects of soil density and vertical macro-pores on the development of cereal crops have been discussed by Ehlers et al. (1983) and Jakobsen and Dexter (1987).

SOIL STRUCTURE AND WATER BALANCE

Infiltration

In order to store water for subsequent use by plants, soil requires a system of pore spaces which are continuous with the soil surface where the water will arrive either by natural rainfall or by irrigation. These pores need to be of sufficient size so that the water can infiltrate and the surface is sloping as fast as it is applied at the surface. If the water arrives faster than it can infiltrate and the surface is sloping, then run-off will occur which will probably cause erosion.

When macropores, such as desiccation cracks, reach the soil surface, run-off from the inter-crack or ped surface can infiltrate the cracks and hence "by-pass" much of the soil matrix (Hoogmoed and Bouma, 1980; Leeds-Harrison et al., 1986). In this case, most of the wetting of the soil peds may be from vertical crack or ped faces rather than from the soil surface. Similar effects can occur with vertical earthworm tunnels (Smettem, 1986). Water which "by-passes" the soil matrix to depths below the depth of rooting in this way will not be available for plant use. This rapid transport through the root zone of the water may also transport fertilizers, herbicides, pesticides, micro-organisms, etc. almost directly into the groundwater (White, 1985).

The ability of soil to transmit water will be impaired severely if a crust forms. Rapid wetting of aggregates at the soil surface can cause them to slake into micro-aggregates of 20–250 μm diameter, and the pores between them may not be able to transmit water at a rate equal to the rate of rainfall or irrigation. Additionally, any dispersed clay may effectively block the pores between the micro-aggregates and give an extremely low infiltration rate. Less slaking occurs with slower rates of wetting of the soil, and when the initial water content of the soil is greater. Less dispersion occurs when the soil pH is lower, and when the electrolyte concentration of the applied water is greater.

Dispersion (or swelling) of clay results in the elimination of the larger soil pores with a consequent significant reduction in soil hydraulic conductivity. Dispersion can be prevented by maintaining a certain concentration of electrolytes in irrigation water. This principle has been applied in the reclamation of saline soils by leaching. Maintenance of electrolyte concentrations above critical values keeps the clay flocculated and this keeps the hydraulic conductivity high (Quirk and Schofield, 1955). Some criteria for maintaining flocculation are shown in Fig. 3. Soil reclamation by the leaching of salt can be done much more rapidly and efficiently if the clay is kept flocculated (Reeve and Bower, 1960).

Infiltration may be severely impeded by soil layers which have been compacted, and also possibly smeared, by tractor wheels and tillage implements. Compaction and smearing, therefore, can result in water-logging, anaerobic conditions, surface run-off and erosion. In rice fields, there is a compacted

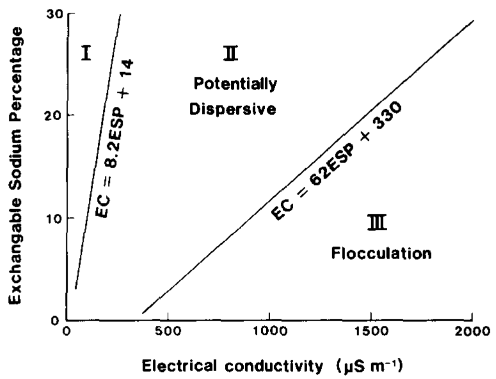

Fig. 3. Flocculation/dispersion diagram for a Red-brown earth in terms of the soil Exchangeable Sodium Percentage (ESP) and the Electrical Conductivity (EC) of the soil solution (1:5 soil:solution suspension). EC is a measure of the electrolyte concentration. In Region I, the soil clay disperses spontaneously. In Region II, the clay will disperse only if the soil is disturbed by moulding (shearing). In Region III, the clay remains flocculated (after Rengasamy et al., 1984a).

layer at the bottom of the puddled layer which may help to reduce water percolation (Mandal, 1984) but this is not a requirement in agriculture in temperate regions. Quite small amounts of compaction can result in large decreases in hydraulic conductivity of the soil (e.g. Dawidowski and Koolen, 1987). Slots filled with stabilized, porous soil in the otherwise dense, sodic subsoils have been tested in Australia as a way to get more rapid and deeper water penetration (Jayawardane and Blackwell, 1986).

Drainage

Drainage of land is desirable in many parts of the world. In wet climates, rainfall can exceed potential evaporation for a large part of the year, and field drainage can prevent water-logging and anaerobic conditions in the soil. After saturation, the soil should return to field capacity in the shortest possible time. This increases the "work days" or number of days when the soil can support traffic and when tillage operations can be carried out.

In dry climates, irrigation practices can make drainage essential. In these climates, irrigation often has to be done with saline water, and excess water (over and above crop requirement) has to be applied to provide leaching. Leaching is necessary to prevent excessive build-up of salt towards the bottom of the root zone. In several areas in Australia, for example, irrigation has been practised sometimes without leaching, sometimes without drainage and sometimes without either. In some of these areas, increasing salinity and/or rising water tables have eventually resulted in abandonment of the land. Obviously, irrigation should not be carried out without adequate provision for removal of any excess of leaching water.

There has been considerable interest in recent years in the development of cheap drainage systems such as mole-ploughing. Mole drains are best made when the soil is drier than the plastic limit. Under these conditions, an array of cracks is formed which runs upwards from the mole tunnel (Russell, 1973; Davies et al., 1979). These cracks are essential for rapid transmission of water to the drain. Any attempt to mole-plough or to do any sort of sub-soil tillage in wet conditions usually does more harm than good. The disturbed soil around mole drain tunnels tends to be rather unstable, and tunnel collapse in some soils and conditions can render the tunnels useless in just a few days (Spoor et al., 1982a,b). However, in many soils moling can be very effective and the drains can operate efficiently for several years.

Deterioration of soil structure by any mechanism usually results in loss of the larger pores and perhaps also loss of continuity of medium-sized pores. This leads to a greatly reduced soil hydraulic conductivity and consequent decrease in drainage efficiency. Soil must be permeable for efficient drainage.

Evaporation

Evaporation loss of water from untilled soil depends on capillary rise to the surface. If the continuity of the capillary pores is lost, then evaporation will be reduced. This may be achieved by tillage to break the capillary pores by producing a "dust mulch" or bed of fine aggregates at the soil surface. Capillarity then can occur only through points of contact between aggregates instead of through a continuous soil matrix. Evaporation is therefore very much reduced.

The pores between the aggregates must not exceed about 5 mm diameter or there will be a considerable convective flow of atmospheric air through the mulch (Farrell et al., 1966; Kimball and Lemon, 1971). Convective flow can accelerate drying of the mulch layer and can also remove significant amounts of water from the surface of the untilled layer beneath the mulch. So, a mulch needs to be fairly fine. Because small aggregates can be readily eroded by water and wind, larger aggregates may need to overlie the fine ones. Some effects of tillage on evaporative water loss are shown in Fig. 4.

SOIL STRUCTURE AND WORKABILITY

The term "workability" is here defined as "the ease of working a well-drained soil to produce a seedbed". This definition excludes the concept of work-days, which has already been discussed in the section on drainage. Everyone involved in tillage knows that different soils differ considerably in workability. The extremes of workability range from the "self-mulching" soils which automatically form a seedbed after one or two wetting and drying cycles (see, for example, Emerson, 1977; McGarity et al., 1984) and the clod-forming soils of

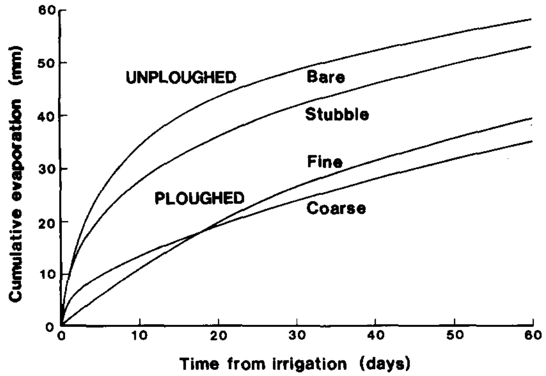

Fig. 4. Effects of soil structure induced by tillage and of stubble retention on cumulative water loss from a South Australian Red-brown earth. The initially dry soil was irrigated with 90 mm of water in the field. Pan evaporation was 9 mm day^{-1} (after Greacen and Hignett, 1976).

Israel with which it is almost impossible to form a seedbed using existing technology (Hadas and Wolf, 1983). Even the same soil can have very different workabilities at different water contents, from year to year, and after different crops. Compared with these environmental effects, the differences in seedbed conditions produced by different tillage implements are small indeed. It would seem, therefore, that the most efficient way to maximize workability is to optimize those soil properties which give rise to workability.

Optimum workability occurs, at least for light- and medium-textured soils, at a water content at or slightly below the lower plastic limit, PL. At this water content, tillage produces the maximum number of small aggregates (Ojeniyi and Dexter, 1979). Field drainage speeds up the removal of excess water and hence speeds up the "return to field capacity". An important factor pointed out by Boekel (1959) is the relationship between field capacity (FC), measured at 100 cm suction, and PL. For soils where the ratio PL/FC is minimum (e.g. 0.55), the soils will drain to a water content where they will behave in a plastic manner on tillage and will not crumble readily. For soils where PL/FC is maximum (e.g. 0.95), the soil will drain to a water content where it will tend to crumble readily. The ratio PL/FC can be maximized and soil structure improved by increasing the organic matter content, the critical level of which depends on the lime status of the soil (Boekel, 1965).

Even if the soil is at FC and PL/FC is maximum, then there is still no guarantee that the soil will be readily workable. Workability requires the soil to have certain structural characteristics. In particular, the mass of soil which is desired to work must contain structure with a mean spacing between structural elements which is similar to the size scale of the aggregates which are required. This structure, for example in the form of a system of micro-cracks, means that there are planes of weakness within the soil. The micro-cracks often define the boundaries between potential soil aggregates. These micro-cracks will

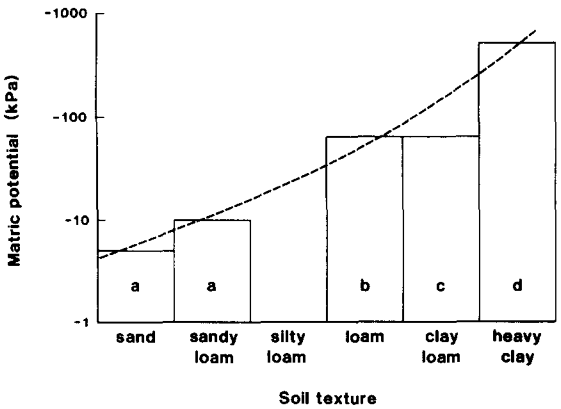

Fig. 5. Upper (wet) limits of workability for soils of different texture expressed in terms of the matric potential of the soil water. Results of: a, Buitendijk (1985); b, Koenigs (1976); c, Boekel (1979); d, Heinonen and Pohjanheimo (1962).

just be air-filled at the optimum water content for tillage. For this reason, the upper (wet) limit of soil workability varies with soil texture as shown in Fig. 5. In a sense, the aggregates may be "pre-existing" or "imprisoned" within the large clods or peds. When a tillage implement comes along, it may impact with the clod or ped which may break along the planes of weakness and hence "liberate" the "imprisoned" aggregates.

The necessary microstructure for workability is often induced by weathering processes such as wetting/drying cycles or freezing/thawing cycles. Russell (1957) went so far as to say that one of the main purposes of primary tillage is to put the soil into such a condition that the weather can act on it readily. The beneficial effects of weathering on workability have been noted by Fountaine et al. (1956), Utomo and Dexter (1981a) and Heinonen (1986), amongst others.

SOIL STRUCTURE AS A HIERARCHY

The previous section introduced the concept of "aggregates within clods". This concept is very powerful and can usefully be extended to other size scales. This idea has been developed by Hadas (1987) who talks about the hierarchical order of soil aggregation. The lowest hierarchical order is the combination of single mineral particles, such as clay plates, into a basic type of compound particle, such as a floccule or domain of clay plates. The next hierarchical order is larger compound particles such as clusters of domains. The next hierarchical order is when a number of clusters are combined into microaggregates, and so on. This hierarchy is illustrated in Fig. 6. Not all of these hierarchical orders exist in all soils.

Compound particles of lower hierarchical order are more dense than those

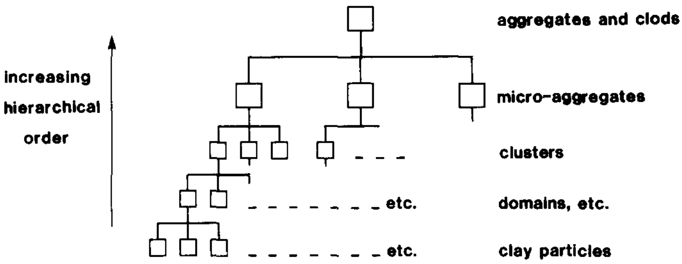

Fig. 6. Hierarchical organization of soil particles or structural elements. There are typically $10^{3\pm1}$ particles of a given hierarchical order in a single particle of the next higher hierarchical order.

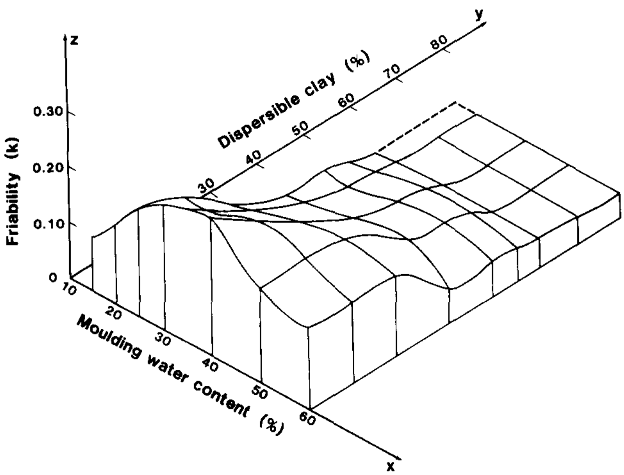

Fig. 7. Soil friability surface showing variation of friability with moulding water content and dispersible clay. Maximum friability occurs when there is zero dispersible clay and when the soil is moulded at or close to the lower Plastic Limit (after Shanmuganathan and Oades, 1982a).

of higher hierarchical order. This is because each order excludes the pore spaces between the particles of the next higher order. This "porosity exclusion principle" is similar to that given by Currie (1966). Compound particles of lower hierarchical order also have a higher internal strength than particles of higher hierarchical order. The strength of the bonds between compound particles of any hierarchical order may be measured by the strength (e.g. tensile strength) of compound particles of the next hierarchical order.

These concepts are compatible with the observation that aggregate tensile strength decreases with increasing aggregate size in a given soil (Braunack et al., 1979; Hadas, 1987). The degree of this dependence has been developed as a measure of soil friability (Utomo and Dexter, 1981b). Friability, k, may be quantified by the slope of a plot of logarithm of aggregate tensile strength against the logarithm of aggregate volume. k is a measure of the scaling law of soil tensile strength.

Effects of dispersible clay and moulding water content on friability are shown in Fig. 7 (after Shanmuganathan and Oades, 1982a). Peak friability occurs at about 38% water content, which is equal to the PL of this soil (60% clay). Dispersible clay blocks micro-cracks thus rendering them inactive and hence reducing friability.

Fractals have recently found increasing application in studies of the scaling laws and of the spatial heterogeneity of a number of soil physical properties (e.g. Burrough, 1983a,b; Armstrong, 1986). It will be recalled that spatial heterogeneity was adopted as the definition of soil structure at the beginning of this paper.

CREATION OF SOIL STRUCTURE

Soil structural features, of a given size order, may be produced either by the combination of structural elements of lower hierarchical order or by the fragmentation of structural elements of higher hierarchical order. These processes are quite different and will be considered separately.

Combination processes

Primary (clay) particles may be combined into quasi-crystals; domains or assemblages depending on the clay particle morphology as described by Rengasamy et al. (1984a) and Oades (1986). Each of these compound particles may be typically 1–2 μm across. These are readily seen with the electron microscope (e.g. Eswaran, 1983).

In a quasi-crystal, thin flexible plates of, for example, Ca-montmorillonite are in almost perfectly aligned stacks with, typically, 80% of the surface areas of individual lamellae in close adhesive contact. The stacks are flexible and may bend as the soil shrinks and swells.

In a domain, rigid platy particles of, for example, Ca-illite have typically 20% of the surfaces of the particles in close contact.

In an assemblage of relatively large blocky particles of, for example, kaolinite, little or no mutual orientation of adjacent particles occurs, and less than 10% of the particle surface areas are in mutual contact. Such groupings of kaolin particles have been discussed by Dickson and Smart (1978). The forces between particles in close adhesive contact and when there are intervening water layers have been reviewed by Quirk (1978), Rengasamy et al. (1984a) and Murray and Quirk (1987).

Compound particles in the size range 2–20 μm are difficult to discuss because there is very little information available on them. Clusters of primary particles can sometimes form compound particles in this size range directly. This can occur by flocculation processes where attractive forces of physical/chemical origin can pull the particles together. Alternatively, clusters of quasi-crystals,

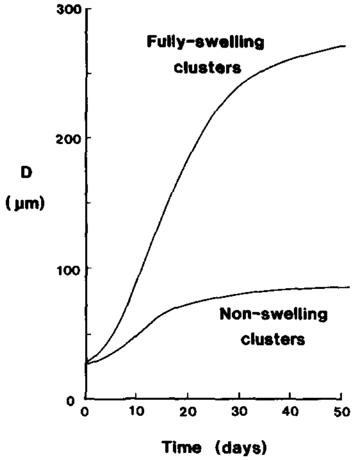

Fig. 8. Spontaneous development of microaggregates in thoroughly-moulded soil as predicted by the theory of Dexter and Spoor (1987) using the tensiometer results and observations of Horn and Dexter (1987a). It still remaines to compare measured and predicted microaggregate diameters and hence to distinguish between swelling and non-swelling clusters.

domains or assemblages may occur. Clusters of 2–20 μm can also be formed by the sticking of clay particles to droplets of mucilage exuded from soil microorganisms or to fragments of partially-decomposed biomaterial. The resulting clusters can be very stable (Turchenek and Oades, 1978).

Microaggregates are conveniently assigned to the size range 20–250 μm. Trapnell et al. (1986) present some excellent photographs and descriptions of microaggregates. Recently, R. Horn and A.R. Dexter (unpublished data, 1988) have observed the formation of microaggregates in several different soils. An example of their results is shown in Fig. 8. Three conditions were proposed for microaggregation to proceed. Firstly, any pre-existing bonds between the soil particles must be broken by mechanical disturbance; secondly, the physical chemistry of the soil must be such that clusters of particles of 10–20 μm diameter can form, and thirdly, the water potential of the soil must be such that water menisci can act on these clusters and can pull them together to form microaggregates with sizes up to the order of 150 μm. Some additional experimental and analytical work by A.R. Dexter and G. Spoor (unpublished data, 1988) has confirmed that the surface tension forces in water menisci are indeed great enough to pull clusters together into microaggregates.

Aggregates, defined as compound particles >250 μm, may be formed in homogeneous soil by wetting and drying cycles. When the soil colloids shrink, cracks have to appear and these cracks define the boundaries of the aggregates or peds. With each wetting of the soil, swelling pressures tend to consolidate the aggregates, but the cracks which defined the boundaries between aggregates always remain planes of weakness. Subsequent wetting and drying of each aggregate occurs through the surface defined by the cracks. This is accompanied by a translocation of clay towards the outer part of the aggregate. This leaves the outer part with a clay content higher than that of the mean for the aggregate, and a porosity which is less than the mean for the aggregate. Conversely, the inner part of each aggregate is more porous and has a smaller clay content than the mean for the aggregate. On each wetting and drying cycle, the features which define the aggregates, such as inter-aggregate cracks and intra-aggregate heterogeneity, become more pronounced until, after a number of years, an equilibrium level of natural aggregation is developed (R. Horn and A.R. Dexter, unpublished data, 1988). Several workers have shown that a substantial proportion of the soil biomass lives on or near the surfaces of aggregates, which again demonstrates spatial heterogeneity.

Clods (> 25 mm) are, in many soils, the result of compaction by agricultural machinery. The compaction pressure applied can press smaller structural units together to form larger stable groupings. Also, the peds delineated by desiccation cracks in a shrinking soil, such as a sodic clay, which contains no significant microstructure may be considered to be clods.

Biological processes, such as the excretion of soil by earthworms, can also produce soil aggregates. It has been estimated by McKenzie and Dexter (1987), that the earthworm *Aporrectodea rosea*, for example, moulds ingested soil in its gut at an extremely low confining pressure of approximately 260 Pa. Thin sections of the casts show little evidence of microstructure. Biological processes cause a wide variety of changes to soil structure. These will not be discussed further here.

Fragmentation processes

Fragmentation of larger structural elements (of higher hierarchical order) to produce smaller structural elements (of lower hierarchical order) is a result of mechanical stress. The stress may be applied externally by, for example, a tillage implement or may be applied internally by the action of water. Either externally- or internally-applied stress will cause fragmentation when the stress in the soil reaches a level of stress equal to the soil strength. The fracture may be either by shear or by tensile soil failure. Tensile failure is more efficient in the sense that less energy is required to produce a given new area of surface of the soil compound particles. Tensile failure is probably also more common in the field, and the rest of this discussion will be restricted to mechanisms of fragmentation by tensile failure.

The drying of a wet soil with a high shrink/swell capacity is shown in exaggerated form in Fig. 9. Initially, shrinkage is accommodated by vertical (downwards) movement of the surface. Eventually, tensile stresses, acting in the horizontal direction (Fig. 9A), become equal to the tensile strength of the soil and vertical desiccation cracks occur (Fig. 9B) (Towner, 1987). Slow drying

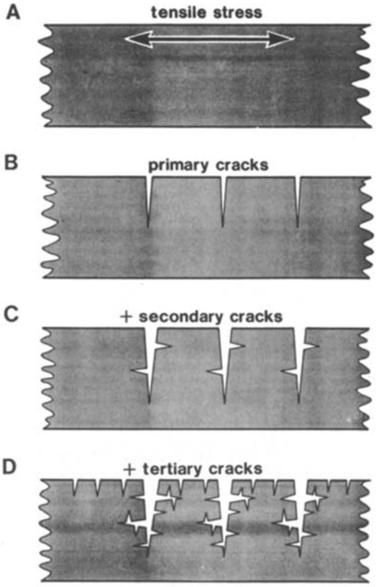

Fig. 9. Development of desiccation cracks in a drying soil as observed on a vertical profile. In A, desiccations producing (horizontal) tensile stresses but cracking has not yet occurred. In B, primary (vertical) desiccation cracks have formed which may form a hexagonal-type pattern on the soil surface delineating peds. In C, drying from ped faces has caused secondary cracking. In D, the soil is as dry as it ever gets and tertiary cracks define peds smaller than in B or C but which are probably still too large for use in a seedbed.

tends to produce wide cracks at large spacings, whereas rapid drying tends to produce more but narrower cracks at smaller spacings. Factors influencing the spacing of tension cracks have been considered by Lachenbruch (1961). These cracks tend to form a hexagonal-type pattern on the soil surface, but in crops they may run half-way between the rows of plants where the soil is wettest and hence weakest (Sharma and Verma, 1977; Nieber, 1983). When the primary cracks reach a certain width (probably around 5 mm), air convection currents develop in them and drying of the soil from the vertical crack faces occurs (Ritchie and Adams, 1974). The tensile stresses induced in the vertical crack walls can then induce secondary (horizontal) cracks (Fig. 9C). Further drying can produce tertiary cracking (Fig. 9D).

The volumes of soil delineated by the cracks would probably be turned up as clods on tillage of such a soil. The clods produced by desiccation cracking of soil are generally too large for use directly in seedbeds, and usually must be further broken down by application of mechanical stresses, externally by tillage, internally by weathering (wetting and drying) processes or by a combination of the two (e.g. a few days weathering followed by mechanical tillage).

Desiccation can bring about fragmentation in another way. Shrinkage of a clay matrix around a relatively large rigid particle, such as a grain of coarse sand, can induce tensile stresses in the clay matrix. The shrinkage of the matrix around a rigid particle is equivalent to the expansion of a cavity (representing the particle) in a matrix. This can produce tensile stresses around individual particles as shown by the theories of Greacen et al. (1968), Vesić (1972) and Misra et al. (1986a). These tensile stresses can cause cracking of the clay matrix either around single sand grains or between pairs of closely-spaced sand grains. Such cracking on drying has been noted by Koolen and Kuipers (1983) and is often associated with the fragmentation of "self-mulching" soils. In some soils it can produce aggregates of seedbed size. Van de Graaff (1978) discusses the possibility of incorporating granules into poorly-structured soils to create centres of stress concentration where soil failure could initiate.

Wetting can produce cracks too. Rapid wetting of a dry soil may cause slaking of the soil into micro-aggregates (Emerson, 1977). Slower rates of wetting can produce partial slaking or mellowing. In this case, the microcracks or incipient failures (in a clod, for example) do not join up to produce completely isolated microaggregates and the soil mass retains its shape and coherence. However, these slower rates of wetting can result in considerable strength reductions (e.g. Dexter et al., 1984), and can increase the soil friability (Utomo and Dexter, 1981b).

The microcracks induced by wetting may not be randomly-oriented, and this can lead to anisotropy of soil strength (McKenzie and Dexter, 1985). Microcracks induced by the stresses generated by differential swelling on wetting are thought to be oriented mainly perpendicular to the wetting front, whereas microcracks induced by entrapped air are thought to lie mainly parallel to the wetting front (McKenzie and Dexter, 1985). Generally, however, it seems that the effects of differential swelling and air entrapment are synergistic in their effects on slaking or mellowing (Grant and Dexter, 1986). Soil will not slake or mellow under the action of wetting unless its initial matric water potential is more negative than about -1 MPa (Sato, 1969; Grant and Dexter, 1986). Rapid wetting does not appear to cause any breakdown of compound particles to sizes smaller than about 250 μm (Oades, 1986).

Mechanical stresses applied externally by, for example, tillage implements or wheels may also cause fragmentation. The theory of the formation of primary (shear) failure surfaces by simple tillage implements is fairly well-developed (e.g. Godwin and Spoor, 1977; Stafford, 1984). However, theories for the subsequent fragmentation or crumbling of the soil as the implement passes by are almost non-existent. Therefore, it is not yet possible to predict accurately the final soil condition which will be produced by a given implement from a given initial soil condition, although there has been an attempt to do this empirically (Dexter, 1979). The principal actions of importance seem to be the impact of the soil elements by the implement which can cause shattering, and the loading of the soil elements by stresses which increase more steadily. In both cases, it is essential to take account of the relative roles of the inertia and the rigidity of the surrounding soil in providing the necessary reaction forces. An example of the effect of loading is the finer tilth left behind tractor wheels running over dry and brittle tilled soil. A simple, preliminary, model for fragmentation in beds of aggregates under uniaxial loading has been presented by Dexter (1987b).

The abilities of roots (Misra et al., 1986b) and earthworms (McKenzie and Dexter, 1988) to crack soil by tensile stresses must also not be overlooked.

STABILITY OF SOIL STRUCTURE

A soil for use in agriculture or horticulture must have not only a structure which is suitable for workability, root growth, etc., but also a structure which will persist for useful periods of time and preferably for ever. The ability of a structure to persist is known as its stability. There are two principal types of stability: firstly, the ability of the soil to retain its structure under the action of water; secondly, the ability of drier (e.g. moist) soil to retain its structure under the action of external mechanical stresses such as the compactive stresses imposed by wheels.

Stability in water

It is usually not very useful to talk about the "destruction of soil structure". What often happens is that one or more of the hierarchical orders may be destroyed, but another hierarchical order may be formed at the same time. An example is the slaking into microaggregates on rapid wetting of dry aggregates which originally contained no incipient failures or microaggregation. In this case, the destruction of a high hierarchical order (the aggregates) is accompanied by the formation of the next lower hierarchical order (the microaggregates). It is necessary to be specific about what has happened.

The destruction of a given hierarchical order automatically destroys all higher hierarchical orders. The ultimate example is the dispersion of the clay particles of a soil in the presence of water. If the organization of the clay particles, being the lowest hierarchical order, is lost, then all the other hierarchical orders are lost or absent.

A good example of this is sodic clay. This has an extremely low hydraulic conductivity (at least with water of low electrolyte concentration), is probably anaerobic, sets hard so that it is impenetrable by roots, and is virtually unworkable. From many points of view, therefore, the most important factor for soil structural stability is to have the clay flocculated. Flocculation is the basis of "good" soil structure.

The conditions for flocculation in terms of the exchangeable sodium percentage on the exchange complex of the clay and the concentration of the electrolyte solution has been shown in Fig. 3. Generally, flocculation is encouraged by a decrease in the thickness of the electrical double layer.

For colloids which have a significant negative mean surface charge density, compression of the double layer may be produced by one or more of the following: increases in electrolyte concentration; increases in the charge of the counter ions so that they reside in the Stern layer; increases in the solution pH at which there is zero net charge on the colloid particles. Some results from Shanmuganathan and Oades (1982b) are shown in Fig. 10. Here, the net charge on the clay particles was modified over a wide range by adsorbing Fe(III) polycations onto them. It can be seen that when the net charge is zero, there is no dispersed clay (i.e. it is all flocculated), and that when the net charge is either positive or negative the particles repel with consequent dispersion.

The stability of higher hierarchical orders in water has been found to depend on different binding agents (Tisdall and Oades, 1982; Oades, 1984, 1986). Some of these are shown in Fig. 11. Micro-aggregates may be bound by polysaccharides which are exuded as mucilages mainly by bacteria, plant roots and fungi. The water stability of micro-aggregates is also enhanced by polyvalent cations (such as Ca^{++}) which can form bridges between the organic colloids and the clay surfaces (Edwards and Bremner, 1967).

Larger aggregates ($>250\ \mu m$) derive much of their water stability from being enmeshed in living or partially-decomposed plant roots and fungal hyphae. The latter are strands of 1–20 μm diameter, may be more than 10 mm long, and may be present in soil in lengths of up to 50 m g^{-1} (Oades, 1986).

The organic matter which stabilizes micro-aggregates is incorporated in the small pore spaces between the domains and other clusters. In these small pores, the organic matter is protected from further microbial attack, and this organic matter may persist for a long time. As shown in Fig. 11, the amount of water-stable microaggregation is rather insensitive to cropping and management. The organic matter which stabilizes the larger aggregates, on the other hand, is constantly renewed by crop growth and is readily accessible to microbial attack. Its amount, and hence the water stability of large aggregates, is therefore sensitive to cropping and management.

These differences are well-illustrated by the mean turnover times for organic carbon in the Rothamsted soil as determined by Jenkinson and Rayner (1977) using radio-isotope techniques. They developed a 5-compartment model in which decomposable plant material, resistant plant material, soil biomass, physically-stabilized organic matter and chemically-stabilized organic matter had halflives of 0.165, 2.31, 1.69. 49.5 and 1980 years, respectively. The proportions of the soil organic carbon in these compartments were 0.04, 1.9, 1.1, 47 and 50%, respectively.

The above observations may be generalized by saying that the organic mat-

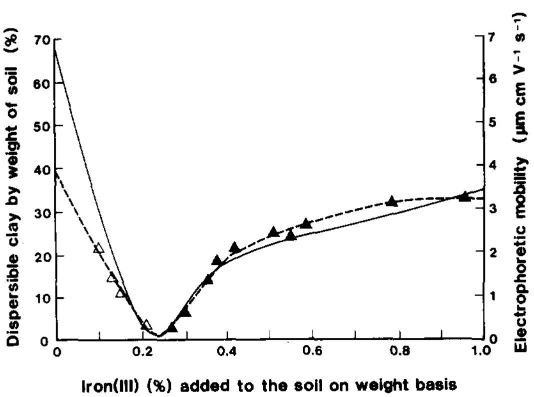

Fig. 10. Dispersible clay in terms of the net charge on the clay particles. Net charge was modified by adsorbing Fe(III) polycations on the clay surfaces and was assessed by measurements of electrophoretic mobility. When the net charge was either positive (▲) or negative (△) clay dispersion occurred. When the net charge was zero, all the clay was flocculated (after Shanmuganathan and Oades, 1982b).

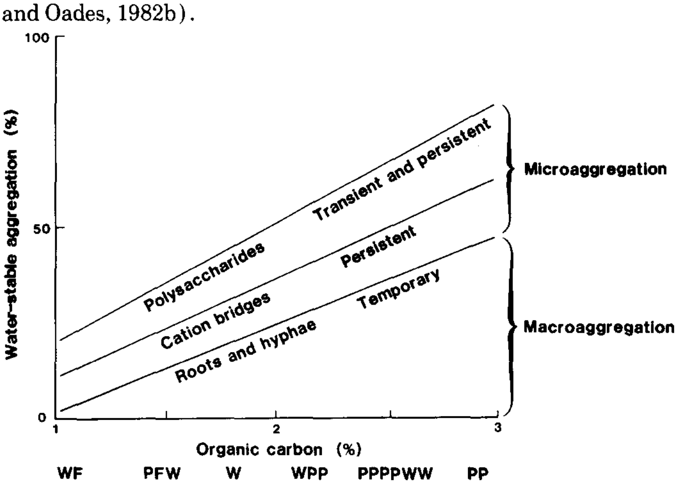

Fig. 11. Water-stable aggregation of a Red-brown earth as influenced by organic-carbon content. The different carbon contents were from plots with different crop rotations as shown where P = pasture, W = wheat and F = fallow. The amount of water-stable aggregates is very sensitive to organic carbon content and hence to cropping and management. However, microaggregates which get their water stability either through polysaccharide mucilages or clay-cation-organic linkages are only slightly influenced by soil organic-carbon content (after Tisdall and Oades, 1982).

ter involved in stabilizing progressively higher hierarchical orders is younger and is increasingly susceptible to soil management practices.

The importance of the disposition rather than the total amount of organic matter was emphasized by Quirk and Williams (1974) who were studying the stabilization of soil by the addition of poly(vinyl alcohol) solutions. They found that maximum stabilization was obtained when the soil was wetted with the stabilizing solution at matric water potentials between −3.3 and −10 kPa. These correspond to equivalent cylindrical pore diameters of 100 and 30 μm, respectively. These pore sizes correspond to those of incipient failures or those between microaggregates. Therefore, stabilization or reinforcement of only this range of pore sizes was necessary to prevent slaking.

Well-aggregated soils often contain substantial populations of both meso- and macro-fauna. But, as pointed out by Oades (1984), it is not yet known whether the fauna are there because of the favourable structure or whether the structure is good because of the fauna. A classic chicken-and-egg problem.

Interactions between soil management and aggregate stability have been reviewed by Quirk (1979).

Mechanical stability

The response of a structured soil to externally-applied mechanical stresses depends on the details of the soil structure and on the stress system applied. By definition, soil will fail when the level of stress exceeds the strength of the soil.

With tensile failure of a compound particle of a given hierarchical order, the failure surface tends to run around and between the particles of the next lower hierarchical order leaving them essentially intact. This is because larger compound particles (of higher hierarchical order) are weaker than smaller compound particles (see Braunack et al., 1979; Utomo and Dexter, 1981b). Therefore, externally-applied tensile stress tends to destroy only the highest hierarchical order present.

With shear failure, the type of behaviour depends on the soil water content, soil density and on the geometry of the applied stress system. Whether shear failure is confined to a shear surface, which may be typically of only 10 particle-diameters thickness, or whether it propagates through the whole soil volume, depends on whether the soil is more or less dense, respectively, than "critical" as described by Critical State Soil Mechanics (Kurtay and Reece, 1970; Hettiaratchi and O'Callaghan, 1980). If a well-defined shear surface is formed, then rolling of round particles or alignment of platy clay particles can occur, but the gross changes occur in only a small proportion of the soil volume. If the shear failure propagates through the whole soil volume, then the whole soil volume may become more dense with the larger pores tending to be destroyed first with consequent reductions in air-filled porosity, hydraulic conductivity, etc.

With compression, densification of the soil can extend through the whole soil volume and through all hierarchical orders as has been shown by the reductions in porosity of clay domains under the action of external pressure (Smart, 1975).

The uniaxial compression of beds of aggregates has been analysed by Dexter

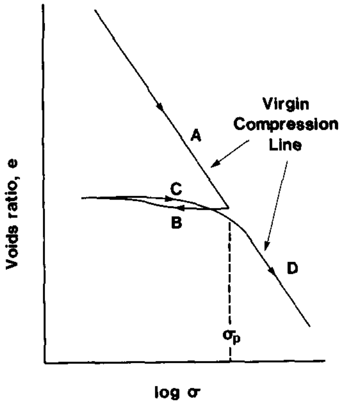

Fig. 12. Virgin compression line (VCL) for a soil at a given water content. Soil compresses (as shown by the decreasing void ratio) linearly and irreversibly with the logarithm of the applied stress, σ (Curve A). If compression is ceased and the stress is reduced, then the unloading Curve B is followed. Further loadings and unloadings at stresses smaller than σ_b will follow the elastic (reversible) Curves B and C. Significant additional compaction will occur only if the applied stress exceeds the "pre-compaction" stress, σ_p, whereupon the VCL will again be followed (Curve D).

(1987b) who expressed the ability of an aggregate bed to support a load in terms of the tensile strength of the individual aggregates in it. Three main types of structural changes have been observed as an aggregate bed is compressed. Firstly, if the aggregates are dry and hard then, under compression, some rearrangement can occur to give packings of lower porosity. These rearrangements are accompanied by small rotations of the aggregates and there is some attrition of the aggregates by friction at their points of contact (Braunack and Dexter, 1978; Guérif, 1982). Secondly, if the aggregates are not so strong but are still subject to brittle failure, then some of the aggregates fracture and the material from these flows into the interstices between the remaining intact aggregates (Dexter, 1975; Guérif, 1982). Thirdly, if the aggregates have a plastic consistency, then the compression will occur by plastic flow with flat areas of contact developing where the aggregates are pressed against each other (Day and Holmgren, 1952; McMurdie and Day, 1958; Davis et al., 1973; Guérif, 1982). For beds of plastic aggregates, Davis et al. (1973) predicted that the inter-aggregate pore spaces would become isolated and cease to be continuous at a value of inter-aggregate porosity of around 5% (v/v). This loss of pore continuity would have a large effect on transport processes in the soil.

Loading of soil will cause compaction only if a certain level of stress called the "pre-compaction stress" is exceeded. In the absence of any age-hardening, this is equal to the maximum previously-applied compaction stress (provided that the water content is the same). This is illustrated in Fig. 12 where the virgin compression line (VCL) for plastic compressive failure is shown. Loading along the VCL (i.e. along Curve A) is irreversible so that removal of the load (e.g. along the unloading Curve B) will leave the soil compacted. Further loadings at stresses smaller than the pre-compaction stress will be along the elastic (reversible) Curve C. Additional compaction will occur only if the pre-compaction stress, σ_p, is exceeded whereupon the VCL will again be followed irreversibly.

Drying of soil can cause compaction through the action of the soil water potential generating effective stresses (Greacen, 1960; Towner, 1961; Childs, 1969; Towner and Childs, 1972; Groenevelt and Kay, 1981; Towner, 1983; Mullins and Panayiotopoulos, 1984; Snyder and Miller, 1985a). Actually, this concept of water potential is oversimplified as described by Passioura (1980), although it is useful for many practical purposes. In this way, drying and subsequent rewetting can leave a soil compacted to a level of pre-compaction stress equal to the maximum level of effective stress applied (Horn, 1981). In a normally-swelling soil, it would be expected that this compaction would be lost on rewetting; whereas, in a non-swelling soil, it would be expected to be irreversible. The level of pre-compaction stress, even of a non-swelling soil, is a function of water content since wet soil is generally weaker than dry soil. Therefore the pre-compaction stress decreases with increasing water content. Soil which is existing at a level of ambient stress below its pre-compaction stress is said to be "over-consolidated" and will dilate upon shearing.

Stress transmission beneath the wheels of agricultural vehicles has been studied by Horn (1983). Generally, the weaker the soil is, the more the load is concentrated beneath the wheel with consequently greater mean levels of compressive stress. A term known as the stress concentration factor describes how the load is concentrated over small areas in the soil beneath a wheel. Therefore, in soil which is weak (either because it is wet or because it has a low density), the concentration factor is large; whereas, in soil which is strong (either because it is dry or because it has a high density), the concentration factor is small. Horn (1983) has studied the effects of soil aggregation on stress transmission. He has shown that the stress concentration factor decreases with increasing level of aggregate development in the order structureless > prismatic > polyhedral aggregation at a given water matric potential.

Soil strength is not a function of only density and water content. This is because of structural changes, such as flocculation and micro-aggregation, which can occur spontaneously with time and because of cementation which can develop between soil particles. These phenomena give rise to thixotropic- or age-hardening of soil after disturbance. The strength of moist agricultural soils which have been completely disturbed by thorough moulding can typically increase by a factor of 2–4 over periods of the order of 10–50 days at constant net density and water content (Utomo and Dexter, 1981c; Dexter et al., 1988).

Age-hardening can occur by two main mechanisms. Firstly, soil particles can rearrange into new positions of minimum free energy. Generally, the lower the free energy, the greater the strength. These rearrangements occur most readily at water contents close to the plastic limit, PL, and do not occur readily outside the range 0.7PL–1.3PL. Secondly, existing bonds at points of particle–particle contact can be reinforced by the diffusion of cementing materials across the particle mineral surfaces. This process occurs at a rate which increases approximately proportionally with the soil water content, and can occur (although very slowly) even in fairly dry soil (-150 MPa water potential, which corresponds to a water film on the soil particle surfaces of one molecular layer thickness) (Kemper and Rosenau, 1984). It is possible readily to identify and distinguish between these two mechanisms through their different age-hardening responses after different levels of compactive stress (Dexter et al., 1988).

MEASUREMENT OF SOIL STRUCTURE

Size and stability of particles

Most soil scientists measure aspects of soil structure in their experiments. The most common is the size distribution of the primary particles and yet very little, if any, use is made of all this data. It is true that good correlations can be obtained between various types of soil behaviour and, for example, clay content; but these correlations are likely only to be applicable to soils having similar clay minerals with similar cations adsorbed on them.

It is also important to query whether it is really the primary mineral particles that one wants to know about. Standard tests for determination of the particle size distribution specify that organic matter and $CaCO_3$ be removed since these cement the clay particles into rather stable compound particles. However, many aspects of soil behaviour in the field such as unsaturated hydraulic conductivity, water retention, soil crusting, erosion and workability are influenced strongly by these compound particles and not directly by the primary particles of which they are composed. It is possible to make a good case for mechanical analysis to be done without removal of cementing agents. Perhaps it is best to do the particle-size analysis both with and without removal of cementing agents and then the differences between the distributions can give some idea of how much of the clay is bound into stable silt- or fine sand-sized clusters or microaggregates.

As has been shown in the previous sections, the compound particles of the various hierarchical orders have a profound influence on soil behaviour in the field. Therefore, an analysis of soil structure should seek to determine what hierarchical orders are present and how stable they are against the actions of water and mechanical stress.

The amounts of water-stable aggregates and microaggregates are readily determined by wet sieving. There are two very different wet-sieving techniques in common use which measure different things. In the first type (as used by Tisdall and Oades (1982) for example), air-dry soil aggregates are immersed directly in distilled water whereupon slaking may occur. A set of oscillating sieves is then used to determine the size distribution of the slaked fragments. Therefore, the proportion of the initial mass remaining on the 250 μm (top) sieve is the proportion of water-stable aggregates, and the proportion passing the 250 μm sieve but being retained on the 50 μm sieve is a measure of the water-stable microaggregation. In the second type of test (as used by Kemper and Koch, 1966), soil aggregates are wetted slowly to saturation in a stream of water vapour and are then immersed in water with no slaking. This test measures the resistance of the immersed soil to breakdown under the action of the mechanical stresses imposed by the oscillating sieves. It is therefore a measure of the strength of saturated soil in the absence of slaking. This test is more appropriate for the extremely fragile soils of the north-western U.S.A. which would always completely slake if wetted rapidly.

As stated earlier, the destruction of a given hierarchical order automatically destroys all higher hierarchical orders present. Clay flocculation, being the opposite of dispersion, is therefore the most basic factor in maintaining soil structure. Emerson (1967) has provided a system for classifying aggregates according to their stability in water. In this system, aggregates are placed in one of seven classes depending on the conditions under which they slake or undergo clay dispersion. This classification is based on relatively simple tests and yet gives a good idea not only of the degree of aggregate stability (or lack of it) but also of the physical or chemical reasons for it. The interpretation of many field experiments would be greatly enhanced if the Emerson dispersion classes of the soils were recorded. It may also be valuable, in many cases, to measure the critical electrical conductivities (or cation concentrations) in the soil solution at which clay dispersion from undisturbed and moulded soil samples ceases (see Fig. 3).

The stability of drier soil is relevant to its ability to support traffic, to be worked, etc. The strength of clods and large aggregates can be measured by the drop–shatter test (Marshall and Quirk, 1950; Ingles, 1963; Hadas, 1984). Basically, this test measures how much soil break-up will occur with a certain energy of impact.It is therefore a relevant test for the assessment of the performance and efficiency of impact-type tillage tools.

For small and/or strong aggregates, the drop–shatter test does not work, and then the crushing test for measuring aggregate tensile strength may be used (Rogowski et al., 1968; Rogowski and Kirkham, 1976; Braunack et al., 1979; Dexter and Kroesbergen, 1985). The crushing test can be used over a wide range of particle sizes and strengths. For irregular aggregate shapes, the test can be calibrated by comparison of the results from plaster-of-Paris replicas of natural aggregates with results from plaster samples of regular shape (Dexter, 1987b) for which theoretical solutions are available. The most suitable regular

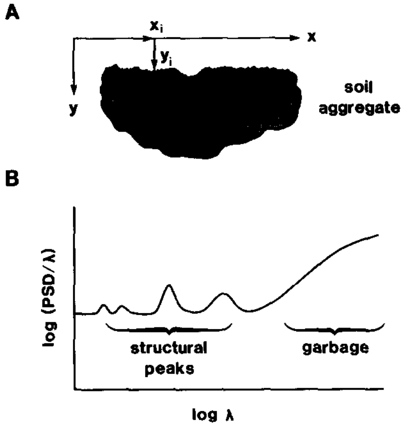

Fig. 13. Fracture Surface Analysis of a soil aggregate. A: the fracture surface is scanned by a video camera and 200–400 points, x_i, y_i, along the surface are recorded by a computer. Spectral analysis of the coordinates of the fracture surface distributes the variance of the y_i values into different component wavelengths, λ. The positions and sizes of the peaks in the power spectrum (PSD) give information about the sizes and spacings of structural features within the aggregate.

shape for the standard in the comparison is the cylinder. The crushing of cylindrical samples, loaded along lines at opposite ends of a diameter, was developed as a test for the tensile strength of concrete (Akazawa, 1943; Carneiro and Barcellos, 1953). Tensile strength may also be measured by direct pulling a sample apart or by a pneumatic fracture method (Snyder and Miller, 1985b). Both of these latter methods are difficult to perform.

A quantitative measure of soil friability can be obtained from the dependence of tensile strength on aggregate size (see Fig. 7 and Utomo and Dexter, 1981b). Friability gives a good idea of soil workability and can be used to assess how factors such as the weather or soil chemical amendments modify soil microstructure. For these reasons, the crushing test appears to be preferable to the drop–shatter test for assessment of the strength and structural condition of soil aggregates.

The sizes, shapes and relative positions of soil compound particles of a given hierarchical order within a compound particle of the next higher hierarchical order may be analysed by fracture surface analysis (Fig. 13). A fracture surface is produced by tensile failure of a compound particle and this is scanned automatically (Fig. 13A). The fracture surface runs around the compound particles of the next lower hierarchical order present. The topography of the fracture surface can be specified exactly by spectral analysis. For methods and applications of spectral analysis see, for examples, Blackman and Tukey (1958); Dexter (1977) and Grant et al. (1985). Peaks occur in the power spectrum of the surface at wavelengths corresponding to the spacings and sizes of surface features (Fig. 13B). Direct observation of the fracture surfaces by scanning electron microscopy can also be most useful. Optical microscopy is not satisfactory as it does not provide the depth of focus necessary to observe rough surfaces.

An important factor in soil management is how soil strength varies with water content. Usually, the logarithm of soil strength increases linearly with decreasing water content. With the Red-brown earth at the Waite Institute, for example, the strength doubles with every 2.5% decrease in the gravimetric water content. In the case of disturbed soil, this dependence is well-measured by the plasticity index (PI) which is the difference in gravimetric water content between the upper (LL) and lower (PL) Atterberg consistency limits. These are best measured by the drop-cone (Campbell, 1975) and rolling-out (British Standard 1377, 1967) methods, respectively. The greater the PI, the more slowly soil strength changes with water content. Although the Atterberg limits relate to freshly-moulded soil, they are very useful and should be recorded whenever possible (especially PL).

Undisturbed soil is generally stronger than disturbed soil and its strength as a function of depth down the soil profile is easily measured with a penetrometer (Anderson et al., 1980). It is true that a penetrometer measures a combination of soil properties and that it cannot be considered as a pure physical measurement method like the triaxial cell apparatus (Bishop and Henkel, 1957). However, soil structural features, such as aggregate size, bulk density and water content, vary rapidly with depth in the field and uniform samples of a suitable size for triaxial testing do not usually exist. It seems logical to use a device, such as a penetrometer, which moves slowly through the soil and senses the strength as it goes. Any method for field use must be rapid and easy so that at least 20 replicate measurements can be made on each of several different soil management treatments within a period of a few hours before the soil water contents change significantly. For this reason, it is also recommended to use drop-cone penetrometers for assessing the strength of surface crusts in the field (Bradford and Grossman, 1982; Campbell and Hunter, 1986) and perhaps Garpenberg probes (Gyldberg, 1986) for assessment of the strength of seedbeds. Because of the sensitivity of soil strength to soil water content, it is essential that accurate determinations of the latter are always made at the same time.

Measurement of the pre-compaction stress (Horn, 1981; Koolen, 1982) gives a good indication of what kind of traffic the soil can support before it is further compacted with consequent loss of the larger pores present. This test should be done on minimally-disturbed soil samples collected from the field in steel cylinders. Again, the water content should always be recorded accurately.

Pore space

The size distribution of soil pores smaller than about 300 μm can be estimated from the drying limb of the water-retention curve using the

Young–Laplace equation. The results are always expressed in terms of equivalent cylindrical diameters or equivalent plate separations. Some uncertainty and error arises from variability of pore size and shape which can lead to constrictions and hysteresis and from lack of knowledge about the contact angle between water and the mineral particles which can be increased by organic coatings on the particles. The greatest problem with this method is soil shrinkage which produces a different total porosity and a different pore size distribution at each level of soil water content. As a result of this, each point on a water-retention curve relates to a different pore size distribution, and it is not possible to estimate a unique pore size distribution at any given water content by this method. Nevertheless, the method can provide very useful information for soils which exhibit limited shrinkage.

The size distribution of pores in initially-dry soils can be measured by a mercury-intrusion porosimeter (e.g. Lawrence, 1977; Lawrence et al., 1979). In this method, mercury is forced into soil at gradually increasing pressures and fills the large pores first. The largest pore size which can be measured is around 150 μm. Positive pressures are required because of the negative contact angle between mercury and soil particles. Samples of clay of high porosity are often very fragile and their structures may collapse in mercury-intrusion porosimetry, thus destroying or changing what it is desired to measure (Lawrence, 1978; Murray and Quirk, 1981).

In order to study the pore-size distribution of a wet clay, either by mercury-intrusion or by direct observation on thin sections or by scanning electron microscopy, it is first necessary to dry the soil with negligible shrinkage. There are three methods for doing this, all of which are based on the concept of drying the soil in the absence of water menisci which give rise to negative water potentials, effective stresses and ultimately to the shrinkage itself. The first is to dry the soil at the critical point at which the menisci disappear (Greene-Kelly, 1973). For water, the critical point occurs under the rather extreme conditions of 22.1 MPa pressure and 647 K. The second method is to freeze-dry the soil. This requires very rapid cooling to below $-150\,^{\circ}$C to minimize the growth of ice crystals which may modify the structure of the sample (Greene-Kelly, 1973). This requirement for rapid cooling throughout the sample places an upper limit on sample thickness of around 1 mm. The growth of ice crystals can be prevented by the prior introduction of dimethyl sulphoxide into the soil water (Keng et al., 1985). The frozen water is removed from the soil by sublimation under vacuum. The third method is replacement of the soil water by organic liquids. The replacement is usually done in the vapour phase by placing the initially wet clay aggregates over the organic liquid (e.g. dioxan or methanol) in a desiccator. This first replacing liquid is then itself exchanged with a volatile, non-polar liquid (e.g. diethyl ether). Finally, this non-polar liquid is removed by evaporation. In a variant of this technique, the replacing liquid may be removed by evaporation at its critical point which may be less extreme than that of water (e.g. 3.6 MPa and 466 K for diethyl ether).

For pores larger than 150–300 μm, therefore, methods based on intrusion or extraction of liquids cannot be used, and recourse has to be made to other methods. These are usually based on the analysis of the sizes and shapes of pore spaces as observed on sections made through impregnated soil samples.

There are two main types of sections in common use: thick sections for analysis by reflected light, and thin sections for analysis by transmitted light. Thick sections are used mainly for the study of structural features larger than about 100 μm whereas thin sections may be used to observe features larger than perhaps 10 μm. With thin sections, it is impossible to get good information about structural features the size of which are smaller than the sample thickness. A commonly-used thickness for thin sections is 30 μm. This thickness can be readily tested because it is the thickness at which quartz (sand) grains change colour under polarized light. The preparation of thin sections is extremely time consuming and requires much specialized equipment especially for the large (150×80 mm) sections of 15 μm thickness which are prepared at Wageningen (Jongerius and Heintzberger, 1975).

The method of impregnation depends on the scale of structural features to be observed and the water content of the soil. To fill pores down to about 500 μm, fairly cheap materials such as paraffin wax (Dexter, 1976), polyethylene glycol (molecular weight 4000) (Willoughby and Walsh, 1972), or plaster-of-Paris (Fitzpatrick et al., 1985) may be used. Plaster-of-Paris can cause problems in shrinking soils by removing water for the water of crystallization which it requires as it hardens. To fill air-filled pores down to 50–100 μm in moist soil, polyester or epoxy resins may be used (Jongerius and Heintzberger, 1975). To fill pores down to perhaps 2 μm for thin sections, the samples should be completely dried, preferably after first replacing the water by an organic solvent such as acetone. The samples should then be impregnated under vacuum, so that entrapped air cannot prevent entry of the resin, and the resin should have its viscosity reduced by the addition of around 50% solvent (e.g. acetone). Some special preparation techniques may be necessary to observe biological features (Tippkötter et al., 1986). For thick sections, resins can have white pigment added to give a good contrast between the pores and the aggregates or microaggregates.

The presence of water in the soil considerably slows down the polymerization of the impregnating medium and hardening can take several months. For the study of the structure of seedbeds in a wet, self-mulching clay, Hewitt and Dexter (1980) completely sealed the epoxy-resin impregnated samples in wax for 2–3 months until the resin had hardened. They then sectioned the samples and measured the aggregate and pore size distributions on the sections before the aggregates (still wet) had dried, shrunk and broken up.

Sections are usually cut with a diamond saw using a non-aqueous lubricant (e.g. kerosene). For coarse seedbeds, wax-impregnated samples can be cut by hand using a normal hand saw. The sections may need to be smoothed and/or polished until the surface roughness (i.e. peak-to-trough height differences) is less than about 0.25 of the size of the smallest features of interest. The softer impregnating materials (e.g. wax or polyethylene glycol) may tend to smear somewhat during the cutting and polishing process.

Porosity can be determined by point-counting in zero-, one-, two- or three-dimensions. Porosity alone is not very useful, however, and what is needed is more detailed information on the sizes, shapes and relative positions of the soil structural features. This information can be obtained by digitizing linear (one-dimensional) transects on soil sections. The transects can be digitized by writing a 1 if there is a particle at a point and 0 if there is a pore at a point. The statistical distribution of 1s and 0s along the transect can then be modelled as a stochastic process (Dexter, 1976; Dexter and Hewitt, 1978; Hewitt and Dexter, 1981). Maximum information about the soil structure is obtained when the spacing between the sampling points is about 0.2 of the size of the structural features of interest. The method can be extended to include more than two different soil structural components and the way that relative frequencies of occurrence vary with their ranking order of occurrence can be used to give the fractal dimension of the soil structure (Hewitt and Dexter, 1984).

This type of analysis of soil structure in one dimension has been particularly useful because plant root axes are themselves essentially one-dimensional. It has therefore been possible to combine the structural parameters of seedbeds, measured in this way, with data on the mechanical properties of root tips and with measurements of soil penetration resistance to provide simple stochastic models for the behaviour of elongating roots in structured soil (Dexter, 1978; Hewitt and Dexter, 1979).

From two-dimensional scanning it is possible to classify the soil pores according to their sizes, shapes and orientations (Murphy et al., 1977a,b; Ringrose-Voase and Bullock, 1984; Ringrose-Voase, 1987). This kind of analysis can be very useful for, for example, investigating size distributions and spacings of desiccation cracks (Guidi et al., 1978), effects of microbial polysaccharides on soil structure (Pagliai et al., 1980), and effects of different soil management practices on soil microstructure (Pagliai et al., 1983).

The study of the true three-dimensional structure of soil is extremely complex and still in its infancy. Although the network of pores in three dimensions can be observed by impregnating them with resin and then washing or dissolving away the soil particles (e.g. Rogaar, 1974), it still remains unclear how three-dimensional pore structures are best measured and described concisely. Lafeber (1965) and Willoughby (1967) analysed the three-dimensional pattern of cracks in soil by analysing a number of successive (two-dimensional) sections. The direction and orientation of each crack was then plotted as a point (or pole) on a polar diagram. After plotting the poles for several hundred cracks in this way, contours of pole density on the polar diagram gave a vivid picture of the statistical distribution of crack directions and orientations in the original soil. More recently, Scott et al. (1986) have obtained similar information from the analysis of crack patterns observed on sections (two-dimensional) taken in several precisely-known directions and orientations. The prediction of structure in three dimensions from measurements made in one or two dimensions is known as stereology (Briarty, 1975; Ringrose-Voase and Nortcliff, 1987). Such predictions have also been attempted using autocorrelation functions (Quiblier, 1984). It seems that there is still much research and development work to be done before we can routinely quantify soil structure in three dimensions with any certainty.

CONCLUDING COMMENTS

There is a growing awareness of the importance of soil structure for all aspects of soil use and management. Soil structure, because of the huge range of size scales involved, affects all of the other properties of soil.

In many parts of the world there is evidence that soil structure and its stability are becoming less favourable for agriculture. The causes include decreasing organic matter levels, compaction, increasing salinity and sodicity, and decreasing levels of soil faunal activity. These changes result in reduced rates of water infiltration and therefore increased water run-off and soil erosion, reduced workability and reduced aeration. In turn, these lead to the more drastic and more obvious economic consequences of increased soil erosion, increased machinery and fuel costs, and reduced crop yields.

Tillage alone cannot remedy a soil which has suffered severe structural degradation. Tillage affects directly only the soil macro-structure. However, tillage can be the first step in soil amelioration by providing air-filled pore spaces in which biological activity can proceed. Increasing the soil organic matter content to increase soil stability takes time: it is not simply a matter of growing a couple of good crops or of ploughing-in a few tonnes ha^{-1} of farm-yard manure. Calculations for the Red-brown earth at the Waite Institute show that it may take 20–40 years for low organic matter contents to be increased towards a new, higher equilibrium content by the growing of continuous pasture. Stabilization of soil requires the proper positioning of the products of microbial decomposition. Soil stabilization was well defined by Quirk (1978) as "the strategy of placing the most appropriate material at the most efficacious place within the soil structure or pore space so that the desired strength may be achieved for agricultural or engineering purposes most economically". Quirk concluded by saying that, in order to achieve this, a more detailed knowledge of the soil pore spaces is required. Here it is necessary to think of pore spaces down to molecular dimensions.

Colloid chemists have developed a reasonable understanding of the conditions under which clays will flocculate or disperse in suspension. Yet the mutual arrangement of particles in wet clays and the morphology of the pore spaces between them remain very little understood. Attempts to measure the pore size distribution of wet clays through the retention or intrusion of liquids tend to destroy the structure which is desired to observe.

At the other end of the scale, our knowledge is little better. The theory which has been developed for soil/tillage machine interactions enables us to predict (and then only for rather simple shapes of implements) the shapes of the primary failure surfaces and the corresponding draft forces. It is not possible to predict the resulting soil condition from any given tillage operation. Too much emphasis has been placed on primary failure surfaces and not enough on the soil crumbling produced by tillage. Increased effort should be directed towards the study of the conditions under which soil fracture and crumbling occur because it is these factors which ultimately control the efficiency of tillage implements. Fracture surface analysis may be useful in enabling us to observe the internal microstructure of clods which is important in the crumbling process.

At intermediate scales (e.g. 2–100 μm), our knowledge is even less satisfactory. This is largely because particles of this scale are too large to have been studied by colloid chemists and yet are too small to be visible to the naked eye. Observations of these particles are best made by low-power scanning electron microscopy on samples which have been dried by vapour exchange of their water with an organic solvent. The conditions under which clusters and microaggregates are formed and destroyed need to be investigated in much more detail. New mechanical tests for measuring the stabilities of these very small compound particles need to be developed. It is the pore spaces between these particles which store much of the water for plant use and which contribute significantly to the unsaturated hydraulic conductivity of the soil.

This review has shown that for a soil to have desirable hydraulic and mechanical properties, and therefore to provide an ideal medium for crop production, it is necessary that all the different hierarchical orders are well developed and are stable against the actions of water and external mechanical stresses. Although we have acquired some knowledge of some parts of the soil structural hierarchy, our overall knowledge is still of a very incomplete and unsatisfactory kind. It is hoped that this review will have identified some of the main gaps in our knowledge and that it may encourage soil scientists of all kinds to explore further into the unknown world of soil structure: our future may depend on it.

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