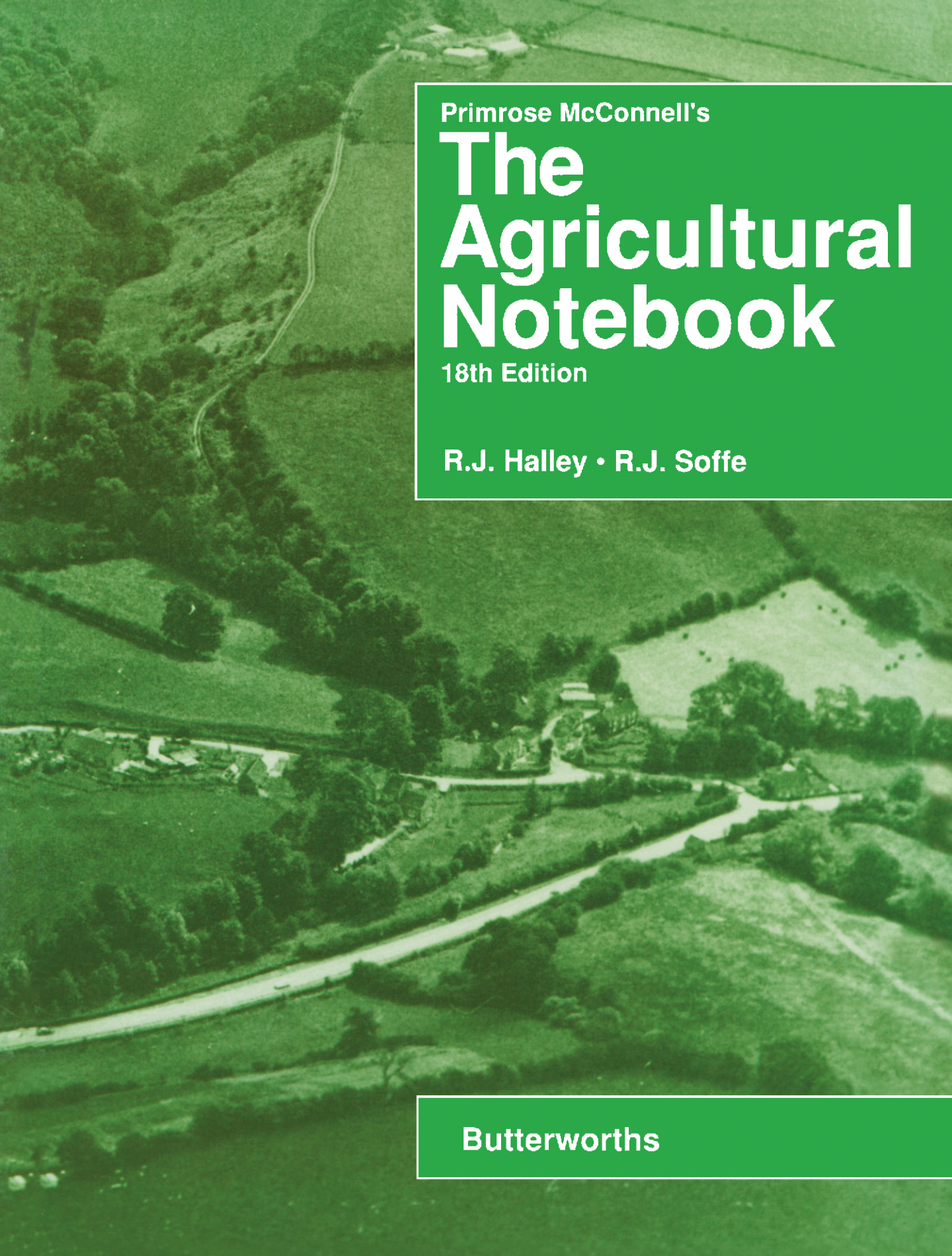


An aerial photograph of a rural landscape, showing a winding road, fields, and clusters of trees. The image is in a monochromatic green color scheme.

Primrose McConnell's

The Agricultural Notebook

18th Edition

R.J. Halley • R.J. Soffe

Butterworths

The Agricultural Notebook

Originally compiled by Primrose McConnell in 1883 this standard work has been completely re-written and enlarged by the contributors listed on p. vii, under the editorship of R. J. Halley and R. J. Soffe.

Primrose McConnell's

The Agricultural Notebook

18th edition

Edited by

R. J. Halley

R. J. Soffe

Seale-Hayne College

Butterworths

London Boston

Singapore Sydney Toronto Wellington

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Preface to 18th edition

The 17th edition introduced a completely new approach and format and this has been popular with readers, particularly students. This new edition builds on that experience by continuing the format and introducing new material to cover developments in the industry over the last six years. It is more difficult in these uncertain times to anticipate in detail needs over the next decade. The Editors believe the current main

UK agricultural enterprises will remain the sheet anchor of the industry, supplemented by new developments of the type covered in the additional chapter on alternative enterprises.

A number of new authors have been recruited and the Editors are grateful to all contributors for their efforts and enthusiasm.

Preface to 17th edition

The first edition of *The Agricultural Notebook* appeared in 1883. It was compiled by Primrose McConnell, a tenant farmer of Ongar Park Hall, Ongar, Essex, who found, when he was an agricultural student under Professor Wilson at Edinburgh University, the great want of a book containing all the data associated with the business of farming. This is the seventeenth edition and all but coincides with the centenary of the publication. The Agricultural Industry has changed so much since the last edition it was decided to abandon the 'pocket book' size and use the new, easy-to-read, format.

Clearly no one person is capable of writing on all the aspects of agriculture necessary for achieving the objectives of *The Agricultural Notebook*. The Editor is most grateful to the team that have contributed to the new edition. Contributors have not been constrained to a common editorial mould but rather, following

discussion with the Editor, have been encouraged to let their own informed aspirations guide the content and layout of their particular sections. Contributors have collaborated and, where appropriate, adequate cross-referencing is given. Compared with the previous edition, some new topic areas have been added and some eliminated to reflect development within the industry and to contain the total volume within an acceptable limit.

The book is aimed at meeting the needs of students of agriculture, be they full-time students in colleges and universities, or practising farmers and advisers. It is hoped the reader will perceive that the essentials of modern science, technology, management and business studies have been carefully blended to give the information which experience suggests is necessary for successful farming today.

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Part 1

Crop production

Soils

A. D. Hughes

WHAT IS A SOIL?

A soil is an ordered combination of minerals, organic matter, air, water and living organisms.

Simple chemical analysis tells us the gross composition of soil but takes no account of the ordered nature of its constituents. Such analysis therefore obscures more useful knowledge than it reveals.

The relative proportions of mineral material, organic matter and 'pore spaces' (which may contain either air or water) vary within very wide limits. For example, peat soils are almost devoid of mineral material, whilst some sands have only very low levels of organic matter. Subsoils usually have lower organic matter levels and fewer pore spaces than topsoils. A useful figure to remember is that in fertile agricultural soils, pore spaces account for about 50% of the total volume. The subdivision of the other 50% between mineral and organic fractions is complicated by the difficulty of knowing the volume occupied by organic matter in soil. By weight (dry soil basis), organic matter levels are typically in the range 2–10% for mineral soils. *Figure 1.1* shows the proportions of the various soil constituents on a percentage volume basis.

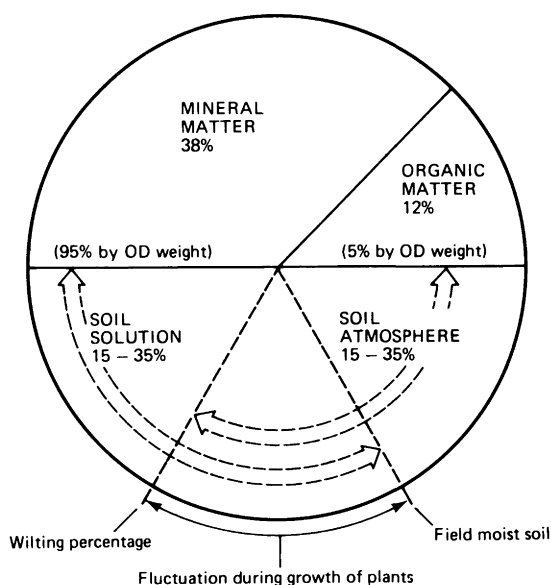


Figure 1.1 Percentage composition of soil on a volume basis (after Jackson, 1964) (from *Chemistry of the Soil*, p. 72, ASC Monograph No. 160, 2nd Edn. Ed. by F. E. Bear. New York: Reinhold, reproduced by courtesy of the author and publishers)

Formation of soils

Soils originate from the rocks of the earth's crust. Rocks are classified as follows:

- (1) *igneous* formed by cooling of 'molten magma', e.g. granite, diorite and gabbro;
- (2) *sedimentary* formed by consolidation of sediments from some previous weathering cycle, e.g. sandstone, shale, dolomite and limestone;
- (3) *metamorphic* formed by action of heat and pressure on sedimentary rock, e.g. slate, marble and quartzite

Note that igneous rocks are formed *de novo* from the earth's molten core whereas material in the other (secondary) types must at some time have existed as igneous rock.

Weathering

Weathering of exposed rock surfaces is due to both physical and chemical agencies.

Physical weathering

Forces of expansion and contraction induced by diurnal temperature variation cause rock shattering, whilst freezing and thawing of any entrapped water greatly accelerates the process. In glaciated areas the major physical agency is crushing and movement by ice and water.

Chemical weathering

Rock minerals are slightly soluble; dissolution of cations and their removal over thousands of years leads to a gradual weakening of the rock's fabric. Chemical weathering is intensified by the acidity resulting from dissolution of atmospheric carbon dioxide, and from organic acids produced during the decay of vegetation. Lichens (which are a symbiotic association of an alga and a fungus) have the ability to fix atmospheric nitrogen as well as being able to fix carbon by photosynthesis. During their growth they accelerate weathering by acidifying the 'soil solution', and rapidly removing the products of weathering as inorganic nutrients. Following their death, carbon is returned to the embryonic soil and provides its first addition of organic matter.

The processes described above can be observed at any time by careful study of the stonework of many ancient buildings and monuments.

Peat soils

A special kind of soil development occurs when very wet conditions prevent the normal breakdown of organic matter by aerobic decomposition. Instead, anaerobic decomposition occurs and the end product is peat. There are two distinct peat formation mechanisms.

Blanket bogs occur in upland areas of the UK where temperatures are low and rainfall exceeds evapotranspiration for at least a part of the year. Heathers and sphagnum moss decay slowly and the organic residues are acid. Any bases (nutrient cations) which may be present are continually leached away by the high rainfall.

Basin peats form in areas where drainage water collects. Native plants are typically sedges and grasses. The incoming drainage water provides ample nutrients so the peat is fertile, having a high base status. Such fen peats are ideal for agricultural cropping when drained, but the act of drainage initiates 'peat wastage', since normal oxidative decay becomes possible. Raised bogs occur when the level of a basin peat surface exceeds the water table height and rainfall is sufficiently great for the blanket bog type vegetation to take over, leading eventually to a blanket bog over the top of fen (sedge) peat.

The soil profile

The vertical section of soil seen in the sides of a pit is the soil profile. Individual layers composing it are called horizons.

Two sets of processes are involved in soil formation:

- (1) physical and chemical weathering giving rise to 'parent material';
- (2) profile development from parent material.

Usually the former precedes the latter but they may proceed simultaneously. Soils where development has proceeded without disturbance exhibit distinctive profiles, their characters being utilised for soil classification and survey.

The upper layer called the A horizon (or 'plough' layer) generally contains appreciable amounts of organic matter which produces a dark colour. This horizon merges into a layer markedly weathered, but comparatively free from organic matter, known as the B horizon. At its base the B horizon, or subsoil, merges into the C horizon or parent

material. The upper part of the C horizon is often considerably weathered and merges gradually into unaltered rock.

THE MINERAL COMPONENTS OF SOIL

The mineral matter in soil is derived by weathering of pre-existing rock – the soil parent material. The mineral fraction of the soil can be subdivided in two principal ways:

- (1) according to particle size; into sand, silt or clay;
- (2) according to mineral composition.

The two are closely related, e.g. fine particles (clay) are composed predominantly of clay minerals, whilst the coarse particles (sand) are normally composed of primary minerals, derived from rock without chemical modification. For practical agricultural purposes in the UK, particle size is the more important attribute.

Sand is largely inert and a very sandy soil is usually well drained, drought prone and lacking in nutrient reserves. Sand grains are usually composed of unaltered primary minerals inherited from the soil parent material; quartz (SiO_2) and feldspars (MAlSi_3O_8 – aluminium silicates of basic cations M^+) are the most abundant of these. The so-called accessory minerals also occur in smaller amounts in the sand fraction. They have a high density and are therefore easily separated from the rest of the sand fraction, and can be used as 'fingerprints' to help identify soil parent material. Weathering of some accessory minerals releases nutrients of importance to agricultural crops. Pyroxenes and amphiboles provide a number of minor nutrients (trace elements) whilst phosphate is released by the weathering of apatite.

Clay is composed of very fine particles with an enormous surface area and great chemical activity. Clay soils hold large amounts of water, much of which may not, however, be available to plants. They are frequently poorly drained, but have good reserves of nutrients.

Silt is insufficiently fine to have significant surface chemical activity, but also lacks the agriculturally attractive properties of sand. Hence silty soils have many of the undesirable properties of both sands and clays and the desirable ones of neither.

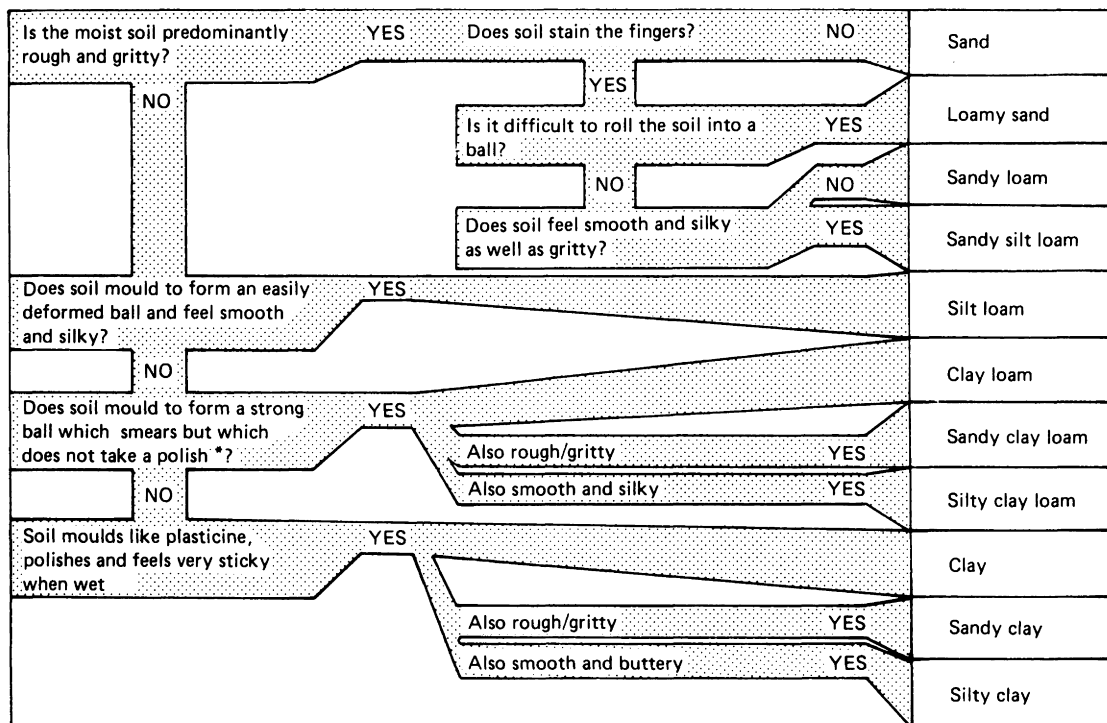
Real soils are always a mixture of sand, silt and clay, with the relative proportions governing soil properties. An ideal soil would have balanced proportions of sand, silt and clay, giving adequate water holding capacity, free drainage, stable structure and a plentiful supply of nutrients.

Soil texture and particle size distribution

The relative proportions of sand, silt and clay determine the texture of the soil. Texture may be derived from particle size distribution obtained by 'mechanical analysis' in the laboratory, or assessed directly in the field by hand.

Hand texturing

For normal agricultural purposes hand texturing is rapid, cheap and reliable. Simply take about a dessertspoonful of soil. If dry, wet up gradually kneading thoroughly between finger and thumb until aggregates are broken down. Enough moisture is needed to hold the soil together and for the soil to exhibit its maximum cohesion:



* Polish = Smooth shiny surfaces to soil when rubbed

Figure 1.2 The assessment of soil texture based on the feel of moist soil

By reference to *Figure 1.2* a texture is assigned to the soil on the basis of its feel.

Particle size distribution (psd)

The process of determining the psd of a soil is referred to as 'mechanical analysis'. The principle is that soils are dispersed in an aqueous solution and separated into size fractions by a combination of sieving (down to about 0.05 mm) and timed sedimentation.

Stokes' equation states

$$V = \frac{g(\sigma - \rho)d^2}{18\eta}$$

where V is the settling velocity in cm/s,
 g is the acceleration due to gravity (about 980 cm/s²)
 σ is the density of the particle,
 ρ is the density of the water
 η is the viscosity of water (about 0.010 at 20°C),
 d is the diameter of the particle (cm).

If an aqueous suspension of soil is thoroughly agitated, then left to stand, the various components will settle at different velocities in accord with Stokes' equation. By carefully sampling from a given depth in the suspension at successive intervals of time, it is possible to calculate the proportion of particles with a given 'equivalent spherical diameter' (ESD). (To make this possible particle density is assumed, by convention, to be 2.6 g/cm³, so that the term $(\sigma - \rho)$ is numerically 1.60.)

Various standard procedures differ in detail, for example:

- (1) pre-treatment – drying and sieving techniques may differ,

- (2) dispersion – use of ultrasonic equipment, treatment with deflocculating chemicals or exchange resins, destruction of organic matter or dissolution of carbonates,
- (3) sedimentation and sampling techniques are fairly standard but an alternative procedure (the Bouyoucos method) uses a hydrometer to measure the density of the suspension at a given depth.

Procedures in common use are described by Russel (1974) and the method used for routine purposes by The Soil Survey of England and Wales and by ADAS is described in MAFF (1986) and Avery and Bascomb (1974). There is no universal agreement on the boundaries of the particle size classes. Systems in common use are illustrated in *Figure 1.3*.

Since numerical percentage composition data are not easily appreciated in speech, it is customary to use a triangular diagram to relate psd data to named textural classes. Again, worldwide standardisation is lacking and a number of systems have been used. In England and Wales both soil surveyors and ADAS advisers use the triangular diagram shown in *Figure 1.4*

CLAY MINERALS

The minerals of the clay fraction may be primary, inherited from the parent material or secondary, formed as a result of chemical processes in the soil. The term 'clay mineral' is conventionally reserved for clay formed as a result of chemical changes in the soil.

The study of clay minerals is a very complex subject and much of the information contained in specialist publications

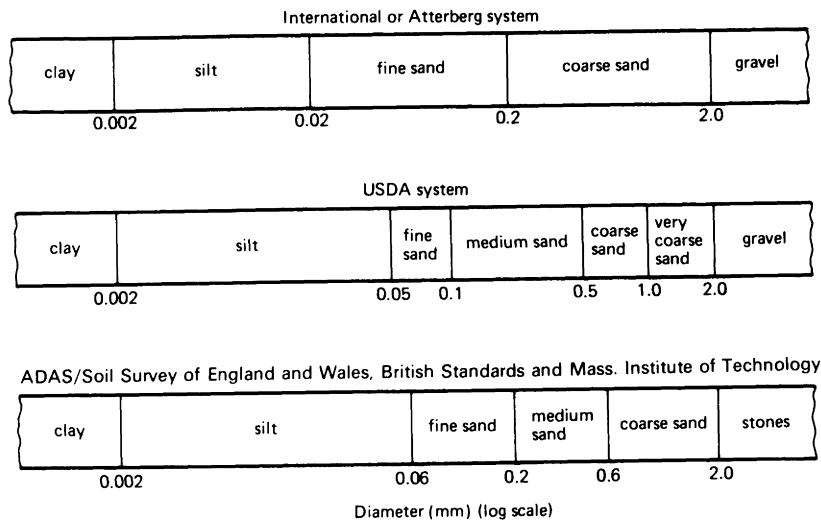


Figure 1.3 Particle size class systems in common use (after White, R. E. (1979) *Introduction to the Principles and Practice of Soil Science*. Oxford: Blackwell Scientific Publications, reproduced by courtesy of the author and publishers)

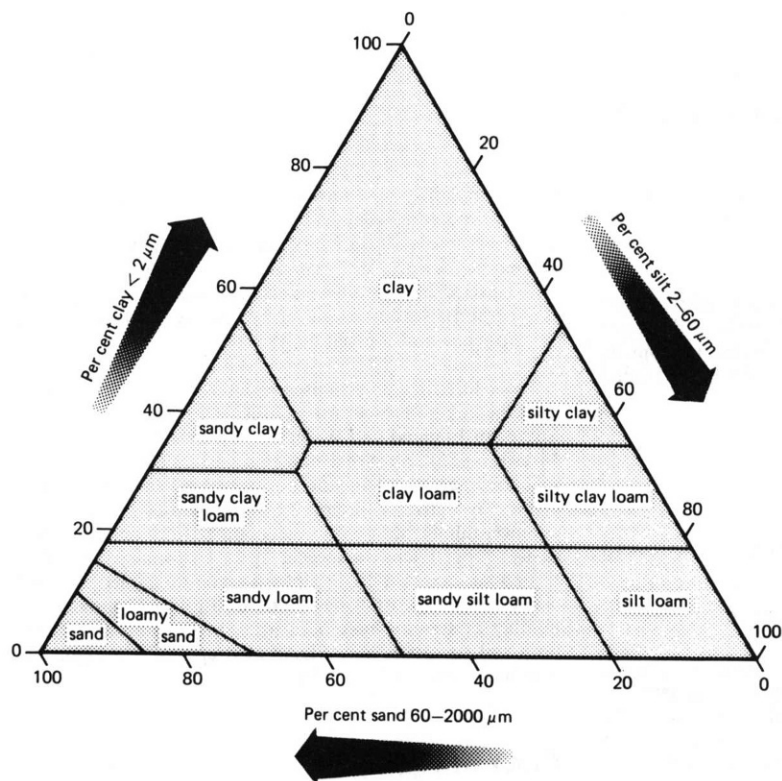


Figure 1.4 Triangular diagram of soil textural classes (after Hodgson, J. M. (1974) *Soil Survey Field Handbook*, p. 23. Soil Survey of England and Wales, Rothamsted Experimental Station, Harpenden, reproduced by courtesy of the author and publishers)

on the subject is of limited practical use to agriculturalists. Agriculturally important properties of soil clays are:

- (1) cation exchange capacity and ability to absorb anions;
- (2) surface area available for chemical reactions;
- (3) ability to shrink and swell in response to changes in moisture content.

Structure of clay minerals

The commonly occurring clay minerals are layer silicates and may be thought of as being built up from planes of oxygen or hydroxyl units. There are two types of oxygen/hydroxyl planes.

- (1) *Complete planes* in which each oxygen touches six others (see Figure 1.5) and some or all the oxygen position may be occupied by hydroxyls.
- (2) *Hexagonal planes* which differ from the 'complete' arrangement (see Figure 1.6) in that (a) there are no hydroxyls, (b) every fourth oxygen position is vacant, (c) the inter-oxygen separation is reduced in the ratio $\sqrt{3}:2$, so that the two types of plane contain the same number of oxygens:hydroxyls per unit area.

Tetrahedral sheets

If a single oxygen is placed on top of a complete plane it will fit comfortably in a position above a space in the plane and beneath it there will be what is called a tetrahedral space (so-called because it is between four spheres whose centres may be thought of as the apices of a tetrahedron). If, instead of one single oxygen, a hexagonal plane is placed above a complete one, a network of tetrahedral holes will occur. In clay minerals these holes contain either silicon (usually) or aluminium cations (occasionally). The two planes and their associated cations are said to form a tetrahedral sheet.

Octahedral sheets

If two complete planes are brought together, a network of octahedral spaces will be formed between three oxygens (or hydroxyls) of one plane and three of the other. A proportion (ranging from 66 to 100%) of these octahedral spaces are occupied by either aluminium or magnesium cations. The two planes and associated cations are said to form an octahedral sheet. If magnesium is the predominant cation all the octahedral spaces are filled and the sheet is said to be tri-octahedral (sometimes called a brucite sheet – brucite is $\text{Mg}(\text{OH})_2$).

If the predominant cation is aluminium, two-thirds of the spaces are filled and the sheet is said to be di-octahedral (sometimes called a gibbsite sheet – gibbsite is $\text{Al}(\text{OH})_3$).

Isomorphous replacement

In the sheet structure just described the arrangement of oxygens, hydroxyls and cations is such that electrical neutrality is maintained. In 'real' clay minerals some cations are replaced by others of similar size but different charge. This replacement does not alter the fundamental structure (form) of the layer but does leave the layer with a net charge.

Thus, aluminium can replace a proportion (up to about

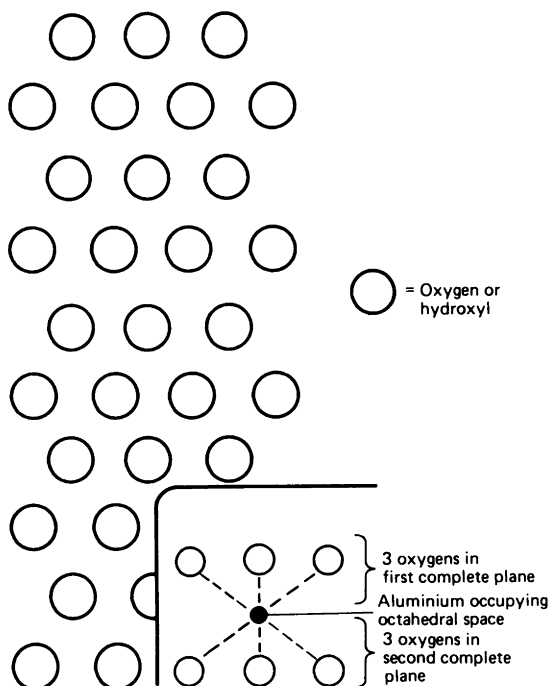


Figure 1.5 Complete plane of oxygens or hydroxyls. Inset – aluminium occupying the octahedral space between oxygens of two complete planes

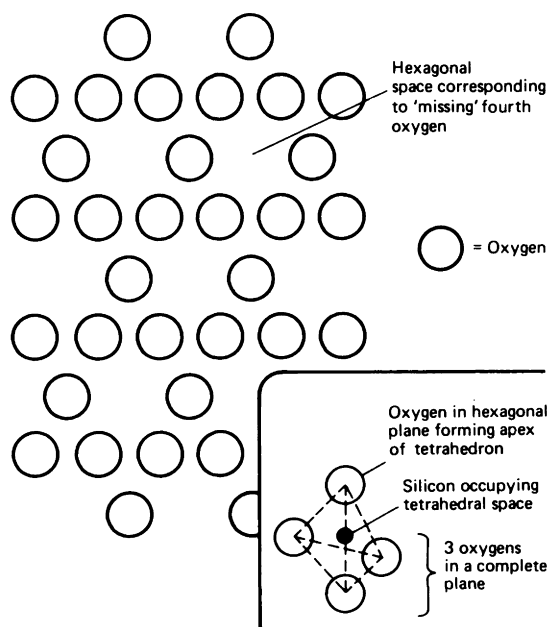


Figure 1.6 Hexagonal plane of oxygens. Inset – silicon occupying the tetrahedral space between oxygens of a complete plane and a hexagonal plane

5%) of silicon in tetrahedral sheets. The larger octahedral spaces can be occupied by a wide range of cations. Isomorphous replacement normally leaves the sheet with a negative charge since every silicon replaced by an aluminium is conceptually the removal of one unit of positive charge, equivalent to a gain of one unit of negative charge. In the same way every octahedral aluminium replaced by a divalent cation contributes one unit of negative charge. Note that isomorphous replacement is a feature of a sheet incorporated at the time of its formation. 'Replacement' implies a departure from an idealised model *not* a dynamic soil process.

Combinations of sheets – layer lattice minerals

The clay minerals may now be visualised as combinations of

tetrahedral and octahedral sheets, some of which may be negatively charged due to isomorphous replacement.

Interlayer bonding

The bonding between adjacent layers may be only oxygen – hydroxyl linkages between uncharged layers which are close together (*see Kaolinite group below*), or by means of cations in the interlayer space neutralising negative charges on the sheets. This type of bonding is readily subdivided into three categories on the basis of the magnitude of the negative charge.

High charge – mica type bonding

The outer sheets of adjacent layers are close together and arranged so that potassium ions can be located where

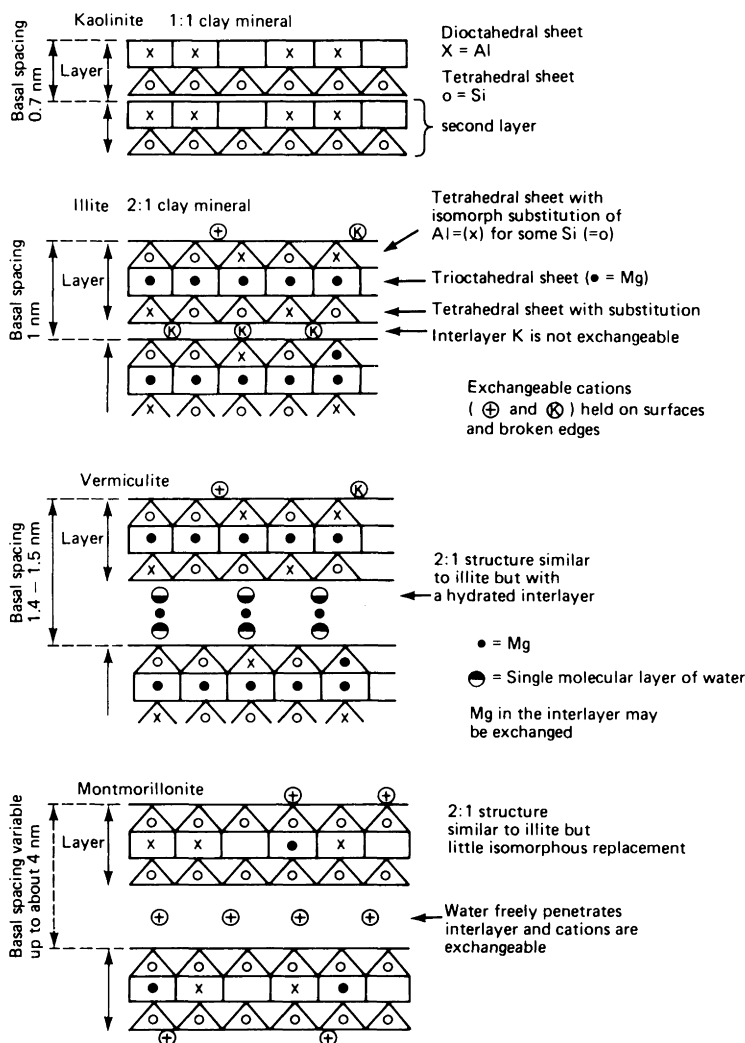


Figure 1.7 Typical layer lattice clay minerals (after Mengel and Kirkby (1982) *Principles of Plant Nutrition*. Bern: International Potash Institute, reproduced with permission of the authors and publishers)

'surface depressions' in adjacent sheets are opposite one another. This gives a strong bond since inter-ionic distances are small and the basal spacing is 1 nm.

Medium charge – vermiculite type bonding

The adjacent layers are further apart than in the mica type and the interlayer space is occupied by two molecular layers of water containing magnesium or aluminium ions. The basal spacing is 1.4–1.5 nm and the larger inter-ionic distance gives a weaker bond, although the interlayer water contributes some hydrogen bonding.

Low charge – montmorillonite or smectite type

The structure is similar to vermiculite, but the bond is weaker and the interlayer cation can be exchanged with solution cations.

Typical layer lattice clay minerals (Figure 1.7)

Kaolinite is known as a 1:1 clay mineral composed of one tetrahedral and one octahedral sheet. There is no interlayer and successive layers are held by oxygen–hydroxyl linkages. Water is not able to move between the units and only exposed surfaces and broken edges contribute to the cation exchange capacity.

Illite is a 2:1 mineral (two tetrahedral sheets sandwiching one octahedral sheet). Adjacent layers are bonded by potassium which forms an integral part of the structure and is not exchangeable.

Vermiculite is similar to illite, but interlayer bonding is due to magnesium (Mg) in association with two monolayers of water. The Mg^{2+} is exchangeable, so intergrades occur and leaching with KCl solution leads to replacement of Mg^{2+} by K^+ and adoption of the illite structure (i.e. 1 nm basal spacing)

Montmorillonite is the commonest of the 'smectite' or 2:1 swelling clays. Lattice structure is again a variation of that of illite; the charge on layers is low (little isomorphous replacement) and bonding is sufficiently weak that interlayer cations are readily exchangeable, and water may enter the interlayer so that there is no unique basal spacing.

Hydrous oxides

Iron (Fe) and aluminium (Al) form hydrous oxides with no recognisable structure (amorphous) and a gross composition represented by $Fe_2O_3 \cdot H_2O$ and $Al_2O_3 \cdot H_2O$. These hydrous oxides occur principally as surface films on clay minerals in soils of temperate lands, and contribute to the cation exchange capacity (CEC) and phosphate fixing capacity. In tropical soils the hydrous oxides may dominate the colloid fraction; phosphate fixation capacity is then extremely large.

Clays in soils

The clay particles present in soils differ from the pure examples already described. In particular, soil clay minerals are formed *in situ* during natural weathering processes, so in all but the oldest and most mature soils intermediate stages of clay mineral formation, as well as end products, are found. In general, soil clays have a less well ordered structure than the pure varieties (usually of geological origin) and both the layers and interlayer ions may differ within a single clay particle. Similarly, adjacent clay particles are not expected to have the same composition, and the picture is further complicated by the ubiquitous occurrence of hydrous oxide films.

Because of this large range of variability, soil clays should be thought of as smectites (montmorillonite type), vermiculite type, illite type or kaolinite type. When successive layers (and interlayers) have different forms and compositions the mineral is said to be interstratified.

Figure 1.8 shows the important features of a soil clay complex formed by weathering of mica.

Clays and soil shrinkage

Laboratory observations show that minerals of the smectite group expand as they are wetted and contract on drying. Casual examination of field soil surfaces or soil profile pits during dry weather reveals the presence of cracks which are entirely absent under wet conditions. Since the cracking

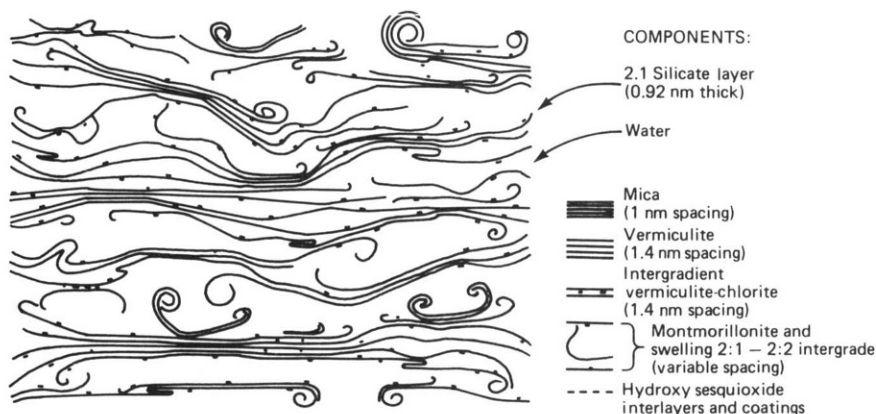


Figure 1.8 Important features of a soil clay complex formed by weathering of mica (from Jackson, M.L. (1964) *Chemistry of the soil*, p. 93. ACS Monograph No. 160. Ed. by F. E. Bear. New York: Reinhold, reproduced with permission of the author and publishers)

observed in the field is most extensive on clay soils, it is tempting to assume that it is due to the same mechanism as the expansion and contraction seen under laboratory conditions. This may be the case in hot climates with a long 'dry season', but under UK conditions field cracking behaviour is not governed by the proportion of smectite in the clay fraction (Reeve *et al.*, 1980). The explanation for this is that the drying experienced in the field is not sufficiently severe to remove the interlayer water removed by laboratory drying. The observed cracking in the field is largely due to 'bulk shrinkage' caused by reduction in the size of soil pores as the soil dries out.

SOIL ORGANIC MATTER AND THE CARBON CYCLE

Soil organic matter (OM) is an extremely emotive topic, associated in many people's minds with 'the goodness of the soil' and man's presumed ill treatment of it. It is also a difficult subject for study. This is because OM is not a single entity – it is a complex system whose individual components are not all capable of precise characterisation. Furthermore, the soil's organic fraction is never static – transformation from one form to another is one of its fundamental characteristics. At any one time, some 2% of the organic matter in an arable soil and up to 4% under grass or woodland are accounted for by living organisms.

The formation of organic matter

Living organisms (plant and animal, including micro-organisms) are the sole precursors of the soil organic fraction. Visible plant remains are not usually considered as 'soil organic matter' but micro-organisms are included, if only for want of any means of isolating them from the soil matrix.

Chemical analysis of organic matter (OM)

In a previous section it was noted that simple chemical analysis of soil obscures more useful knowledge than it reveals. This is particularly true for OM.

Classic techniques for fractionation of OM involve successive treatments with alkali and acid or extraction with ether and other organic solvents. The 'fractions' isolated are not homogeneous and they do not exhibit the recognisable properties of soil organic matter. Furthermore, many fractionations leave the majority of the organic matter in the residue which is resistant to the methods of the analytical organic chemist, and therefore only readily characterised by its total elemental analysis.

Properties and functions of organic matter in the soil

'Chemical' classifications of OM are of limited relevance to agricultural problems. The serious agriculturist should beware of being misled by the great volume of specialist literature relating to them.

Organic matter and soil structure stability

The bonding together of fundamental soil particles to form the structural units which are such an important feature of productive soils is a complex process involving clay mineral particles bonded to each other and to organic matter by chemical forces to form so-called 'microaggregates'. Similar mechanisms together with the mechanical actions of plant roots and fungal hyphae bind the microaggregates together to form larger units.

Polysaccharide and polyuronide gums (mostly of bacterial origin) may be adsorbed onto soil particles. Such gums typically have a molecular weight between 100000 and 400000 so one molecule can be adsorbed onto more than one mineral particle, forming a very strong link.

Additionally, charged groups on organic molecules (e.g. amino groups) can form bonds by electrostatic linking with opposite charged sites on clay minerals and again one large molecule may interact with more than one mineral particle. The so-called 'cation bridge' which involves linking of a negative charged clay site to a negative charged organic anion through a polyvalent cation can be viewed as a variant of this theme.

Exchange capacity

Many of the chemical properties of soils are dependent upon cation exchange capacity (CEC – sometimes called base exchange capacity). Organic substances contain carboxyl groups which can dissociate to provide negatively charged sites. The charge is typically 2–4 mEq/g compared with 0–2 mEq/g for clay minerals. Thus even though the OM content of most soils accounts for a small proportion of the total soil mass, it may make a significant contribution to the CEC.

Mineral nutrition of crops

Since nitrogen, phosphorus and sulphur are important constituents of OM, eventual breakdown by micro-organisms releases quantities of these elements to the soil in a plant available state. (Nitrogen, phosphorus and sulphur cycles are discussed more fully on pp. 20–25.) Some organic substances form complexes with copper, and hence its availability to crops is also influenced by OM formation and breakdown.

Organic matter and agricultural practice

Notwithstanding the complex chemical nature of OM and the mechanisms whereby it affects soil properties, there is no doubt that high OM levels are generally associated with ease of management of soil (*see for example Low, 1972*).

Soil organic matter level

In the laboratory, soil organic matter is normally determined by wet oxidation with a dichromate/sulphuric-phosphoric acid solution. The dichromate remaining after oxidation is determined by titration with ferrous sulphate solution (*see MAFF, 1986*). The method gives a value for percentage organic carbon: by convention this is multiplied by 1.72 to give 'organic matter percentage'.

The oxidation reaction is invariably incomplete since a

Table 1.1 Quantities of organic material returned by roots of various crops (after Davies, Eagle and Finney, 1972)

<i>Crop</i>	<i>kg/ha of dry roots in top 20 cm</i>	<i>% increase in total organic matter in 20 cm of soil before decomposition</i>
1-year grass ley	4500–5500	0.2–0.3
3-year grass ley	6750–9500	0.3–0.5
Winter cereals	2500	0.1
Spring cereals	1450	less than 0.1
Sugar beet	550	less than 0.1
Potatoes	300	less than 0.1
Red clover	2200	0.1
25 tonnes farmyard manure	4500	0.2

From *Soil Management*, Table 27, p. 194 by courtesy of authors and the publishers. Farming Press: Ipswich.

proportion of the OM is protected by intimate association with clay, but the figures obtained are well established in the literature and serve adequately for most comparative purposes.

The OM level in a soil at any given time is the result of a balance between additions and losses over its recent history. The factors affecting the balance are as follows.

Climate

In higher rainfall areas or where drainage is poor, the rate of decomposition is lower and OM levels are higher. Thus for any given soil texture and cropping sequence the OM level will be higher in the west of the UK, and at higher altitudes.

Soil type

Because of the intimate association between humus and clay particles, the OM in clay soils is better protected against attack and degradation than in sandy soils. Also clay soils are in general more likely to remain moist for longer periods of the year. Both these factors lead to clay soils having higher OM levels than lighter soils under otherwise similar conditions.

Cultivation

Frequent cultivation is thought to increase the rate of OM loss. This is not now believed to be a simple effect of better aeration leading to faster oxidative breakdown of organic materials. Powlson (1980) has suggested that the disruptive forces of cultivation result in exposure to micro-organisms, of organic matter which had previously been inaccessible to them. Also some micro-organisms are killed when soil is disturbed and their premature decay contributes to the loss of OM.

Less soil disruption and consequent slower OM breakdown in the surface horizon may be a beneficial effect attributable to 'non-plough' cultural methods.

Cropping

There has been interest in the effects of crop remains returned to the soil since chemists first started to look at agriculture. Typical quantities involved are shown in *Table 1.1*.

The main point illustrated by this table is that the amounts of organic material returned to the soil from crop roots or manure dressings are too small to affect soil OM levels appreciably, even if no adjustment is made in recognition of the fact that the majority of organic residues returned to the soil are lost by respiration of micro-organisms, and only a small proportion is eventually incorporated into humus.

Note also that contrary to common belief, the root remains from a good winter cereal crop makes a sizeable contribution to the soil's OM economy (equivalent to about 14 t/ha of farmyard manure).

Effects of leys on soil OM levels have been studied by a series of experiments – at the ADAS (formerly NAAS) experimental husbandry farms. Results are summarised in *Table 1.2* and show that the increases due to a three-year ley are small and those following nine years of ley are only modest; note also that the gains of nine years' grass were virtually wiped out by a mere three years of arable cropping.

These results are consistent with the view that the structure stabilising powers of leys are due to gums produced by actively growing soil bacteria, the clay binding powers of bacteria themselves, and the mechanical effects of both grass roots and fungal hyphae. All of these agents are ephemeral and their decline following return to arable cropping would be reflected by a fall in total OM level.

Permanent pasture usually has higher OM levels than leys under similar conditions, and at least a proportion of it is of

Table 1.2 Effect of leys on level of organic matter in soils from ley fertility experiments (from Davies, Eagle and Finney, 1972)

<i>Experimental husbandry farm</i>	<i>Location</i>	<i>Soil</i>	<i>% organic matter in topsoils</i>			
			<i>at start of trial</i>	<i>after 3 year ley</i>	<i>after 9 year ley</i>	<i>after 3 year arable following 9 year ley</i>
Boxworth	Cambs	Clay loam	3.1	+0.1	+0.5	+0.0
Bridgets	Hants	Chalk loam	4.4	+0.1	+0.6	+0.1
Gleadthorpe	Notts	Sand	1.6	+0.1	+0.36	+0.3
Rosemaund	Hereford	Silt loam	3.5	+0.2	+1.0	+0.0
High Mowthorpe	Yorks	Chalk loam	3.8	+0.4	+1.0	+0.5

From *Soil Management* Table 28, p. 196 by courtesy of the authors and the publisher Farming Press: Ipswich.

a more durable nature, so that it is lost only slowly following a change to arable cropping.

The carbon nitrogen (C:N) ratio of organic matter of cultivated soils in temperate regions is fairly constant, having a value between 10:1 and 12:1. When manures with wider ratios than this are added to soil, either a disproportionate amount of carbon must be lost by microbial respiration, or additional N taken from some other source to compensate. (This is dealt with in more detail on p. 21.)

The carbon cycle

The transformations of carbon from gaseous carbon dioxide in the atmosphere to plant, animal, and thence soil OM and

finally back to the atmosphere, form a cycle depicted in Figure 1.9.

Carbon dioxide (CO_2) in the atmosphere is the source of carbon for plant growth, and the 0.03% CO_2 in the world's atmosphere represents a total quantity of about 7×10^{11} tonnes of carbon. This is a relatively small amount by comparison with the estimated 9.6×10^{13} tonnes of carbon in fossil fuels (coal and oil) and 12×10^{15} tonnes in sedimentary rock (Ehrlich *et al.*, 1977).

The rate of fixation by photosynthesis depends on climatic conditions, soil and crop. Maximum rates (worldwide) are about 50 g of dry weight/($\text{m}^2 \text{ d}$) (200 kg/(ha d) of carbon). In the UK winter wheat makes the majority of its growth in the period April–July (120 d) and the average rate of increase in

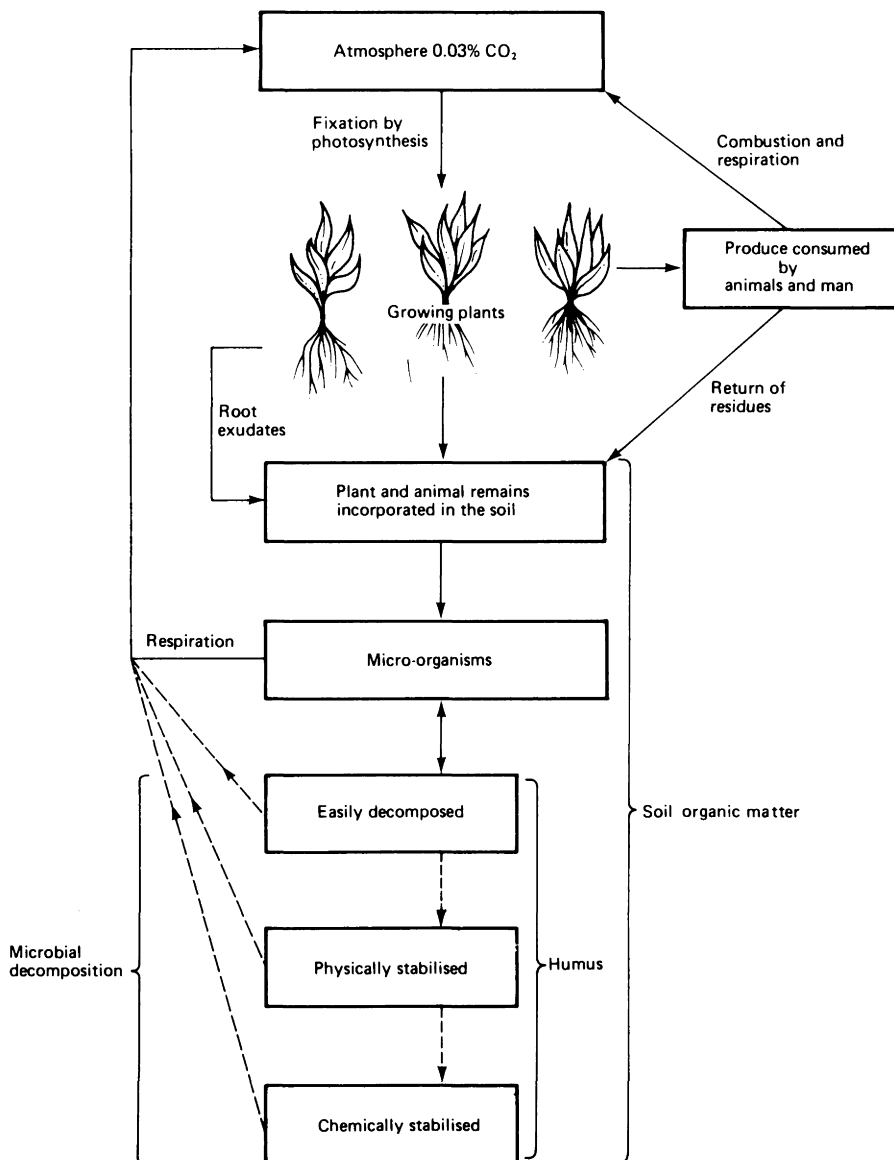


Figure 1.9 The carbon cycle

grain dry weight over this period is 0.1 tonnes/(ha d) or 10 g/(m² d). Making allowance for leaf and root growth, 20 g/(m² d) is a reasonable figure for this crop's total dry matter (DM) production. This corresponds to a carbon fixation rate of 80 kg/(ha d).

Plant and animal remains (including excreta) are the source of carbon from which soil organic matter is formed and quantities returned by roots of various crops have already been quoted (*Table 1.1*). Additionally, root systems contribute to soil organic matter by production of root exudates. These comprise both soluble (sugars, amino acids, etc.) and insoluble compounds (polysaccharides, sloughed-off cells, etc.). They make a direct contribution to the nutrition of micro-organisms in the rhizosphere and are thought to be responsible for the very high level of microbial activity in that zone.

Because of the close association of soil, micro-organisms, roots and their exudates, measuring the quantity exuded has proved extremely difficult. Estimates vary, but 10% of the total above ground dry weight over a season seems a likely order of magnitude.

This input of organic matter to the soil may be visualised as a flux of carbon being 'processed' through a volume of soil according to the equation

$$\frac{dC}{dt} = A - kC$$

This assumes that decomposition is what is known as a 'first order reaction' and k is its 'rate constant'; A is the annual addition of residues and dC/dt is the rate of change of the soil's organic carbon level (by convention organic matter (% OM) is derived from a laboratory determination of %C by multiplication by 1.72). For an equilibrium state $dC/dt = 0$ and the 'turnover time' is given by

$$\frac{C}{A} = \frac{1}{k} \text{ (years)}$$

Annual inputs of carbon for old arable soils at Rothamsted are typically 5% of soil organic carbon, so turnover time is calculated to be 20 years.

Radioactive carbon-14 dating techniques have allowed the mean age of carbon in existing soil OM to be measured, and results are in the region of 2000 years. Closer study has shown that different components of soil OM have differing turnover times.

Easily decomposed humus has a turnover time of about three years.

For physically stabilised humus – that which is protected from easy microbial attack by its location in fine pore spaces – the average figure is about 100 years.

Chemically stabilised 'ligno humus complexes', which may be protected by adsorption onto clay mineral surfaces, have mean turnover times as large as 4000 years.

The population of micro-organisms at any one time accounts for 2–3% of the soil's total organic carbon and this fraction has a turnover time of two to three years.

The 'priming action' of fresh organic matter additions

At one time it was thought that the enhanced microbial activity occurring when fresh OM was added to the soil could lead to accelerated breakdown of the old and more stable humic substances. By use of carbon-14 tracer techniques, Jenkinson (1966) was able to show that this

'priming effect' generally had a negligible effect on the overall rate of decomposition, any losses being compensated for by the biomass of the enlarged population of micro-organisms.

CHEMICALLY ACTIVE SURFACES AND THE CHEMISTRY OF SOIL COLLOIDS

The seat of chemical activity in the soil is the colloid fraction. This comprises the clay minerals and organic matter.

The soil's power to hold (adsorb) cations

Cations, represented as M^+ (monovalent, e.g. Na^+ , K^+) or M^{2+} (divalent, Ca^{2+} , Mg^{2+}) are held by electrostatic attraction to negatively charged sites in the soil colloid fraction.

The negative charge on soil colloids is due to:

- (1) *isomorphous replacement* of trivalent by divalent cations within clay mineral lattices, leading to a deficit of charge;
- (2) *unsatisfied charges* at broken edges of lattices;
- (3) *dissociation of protons*
 - (a) from organic acid groups

$$-COOH \rightarrow COO^- + H^+$$
 - (b) from hydrous oxide surfaces

$$M-OH \rightarrow M-O^- + H^+$$

(M is Si (silicon), Mn (manganese), Fe (iron) or Al (aluminium)).

Note that such dissociations are pH dependent, so that the negative charge (i.e. the CEC) is greater at high pH, and some protonation may occur at low pH to give sites of positive charge capable of interacting with anions (see the section 'The soil's power to hold anions', *below*).

Exchange of cations

Ions held by the electrostatic forces at charged surfaces are also subject to the kinetic forces of thermal motion, and the strength of their retention is the result of the balance between the two. At equilibrium a characteristic distribution pattern described by the Gouy–Chapman model is observed.

The Gouy–Chapman model says, in effect, that distribution of charged species within a solution is distorted near to a negatively charged colloid surface, there being a disproportionate concentration of cations near to the surface. An inner conceptual layer, next to the charged surface, is composed solely of cations (Stern layer) with a gradual transition (the Gouy or diffuse layer), to free solution. As shown in *Figure 1.10*, these two layers have an excess positive charge, balanced by the negative charge on the surface and cations in the 'double layer' are in equilibrium with those in the solution. Increasing the concentration of the external solution compresses the Gouy layer, so that the change from the all-cation Stern layer to free solution occurs over a smaller distance. Changing the composition of the solution leads to corresponding changes in the double layer system. Thus, if a clay is saturated with Ca^{2+} and in equilibrium with a $CaSO_4$ solution addition of a K^+ salt will lead to replacement of some Ca^{2+} by K^+ ($2K^+$ replacing each Ca^{2+}), i.e. cation exchange. Some cations exchange (or form Stern

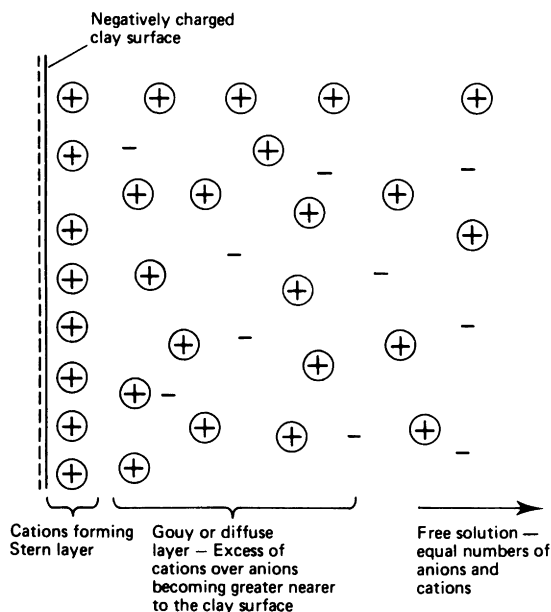


Figure 1.10 The double layer system associated with a negatively charged clay surface

layers) more readily than others. Small, highly charged ions are held preferentially to large ones with low charge, but small, strongly hydrated ions behave similarly to larger ones. Exchange reactions are reversible and the thermodynamically less favourable ones are able to occur, providing the concentration in the external solution is great enough. Thus strongly bound Ca^{2+} will be replaced by less strongly bound K^+ , provided that the K^+ concentration in the external solution is high enough (i.e. a mass action effect).

The *Gapon Equation* is a physical relationship that governs the equilibrium between exchangeable and free cations of two species. For a monovalent cation C^+ and a divalent cation C^{2+} the equation states:

$$\frac{\text{C}^+ \text{ ads}}{\text{C}^{2+} \text{ ads}} = k \frac{a_{\text{C}^+}}{\sqrt{a_{\text{C}^{2+}}}}$$

$\text{C}^+ \text{ ads}$ = adsorbed monovalent cation

$\text{C}^{2+} \text{ ads}$ = adsorbed divalent cation

a_{C^+} = activity of monovalent cation in solution

$a_{\text{C}^{2+}}$ = activity of divalent cation in solution

$\frac{a_{\text{C}^+}}{\sqrt{a_{\text{C}^{2+}}}}$ is called the activity ratio, AR.

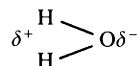
Hence

$$\frac{\text{C}^+ \text{ ads}}{\text{C}^{2+} \text{ ads}} = k \cdot \text{AR}$$

k is primarily dependent upon the strengths of adsorption of the two species.

Adsorption of water

Water is a dipolar molecule with an asymmetrical distribution of charge:



Thus a layer of water molecules may be attracted to a negatively charged colloid surface and they will all be oriented with their δ^+ ends towards the surface. They will thus expose a new surface of δ^- charges which may in turn attract a second layer, and so on. The proximity to the negative colloid surface will increase the effective charge on the innermost layers of water by induction, but this effect decreases with distance from the colloid and so the strength of bonding also decreases towards an outer surface at which electrostatic attraction is balanced by the forces of thermal motion.

The soil's power to hold anions

Positively charged colloid surfaces

Although soil colloids are predominantly negatively charged, localised positive charged sites also occur, particularly on clay mineral edges and on hydrous oxide surfaces.

Of the nutritionally important anions in the soil NO_3^- is not adsorbed to any appreciable extent and sulphate is adsorbed only slightly. Although phosphate is fixed and tightly held by soils, it is incorrect and misleading to think of phosphate being exchanged by negative surfaces (*see below*, 'Ligand exchange'), so overall the positive charge sites on soil colloids are relatively unimportant.

Ligand exchange

Phosphate, silicate, fluoride and molybdate are bound by ligand exchange. This reaction is very important as the mechanism for 'phosphate fixation'. On iron or aluminium hydroxide surfaces, some oxygen atoms are less strongly bound to iron or aluminium than those in the body of the material. They may be replaced by negatively charged species such as phosphate anions. This reaction is characterised by an increase in the negative charge on the surface. It is not reversible and the exchange species can only be displaced by another anion capable of further increasing the negative charge.

Quantity intensity relationships

Although roots take up nutrients from the soil solution, as already explained, the solution is in equilibrium with a solid phase capable of replenishing losses. Thus nutrient availability to plants depends not only upon concentration in the solution, but also on the ability of the solid phase to sustain that concentration against depletion. The term Quantity (Q) refers to the amount of a nutrient, and Intensity (I) reflects the strength of the retention whereby it is held. Concentration in the equilibrium solution is a convenient measure of intensity, since if a species is weakly adsorbed it will support a high concentration in the equilibrium solution. *Figure 1.11* shows two contrasting Q/I relationships, or isotherms.

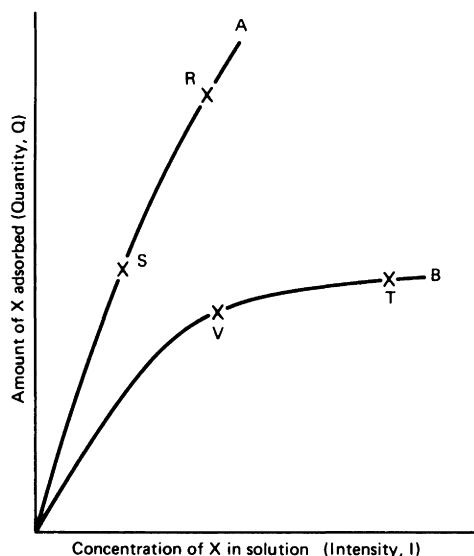


Figure 1.11 Quantity/Intensity relationships for adsorption of 'X' by two contrasting soils

Going from R to S along A shows that the soil is able to sustain the intensity of X as uptake proceeds and the quantity (Q) in the reserve is reduced. From T to V along B a small uptake (reduction in Quantity) leads to a large fall in intensity and the soil is said to be 'weakly buffered with respect to X'.

Flocculation and deflocculation of clays

Clays have properties that enable their particles to either repel one another, or to attract one another and form stable groupings called 'flocs'. Flocculation of clay is vitally important in agricultural soils and fortunately deflocculation rarely occurs in the clay of UK soils.

The mechanism of flocculation

A clay mineral may be visualised as a lattice structure carrying negative charges. It has already been pointed out that positive charges may occur on broken edges. If two such particles are brought together edge to face they will attract one another and many such would flocculate into larger units (see Figure 1.12). This mechanism is responsible for the structural stability of clay soils.

If anything happens to suppress the positive charges, flocculation is prevented. Consider, for example, a clay saturated with Na^+ ions in a strong NaCl solution (e.g. flooded with sea water). A Gouy–Chapman double layer will form where there are negative surface charges, and it will be compact due to the concentrated solution (see Figure 1.13a).

If the external solution becomes less concentrated (e.g. rainfall, or irrigation after flooding) the double layer expands and the clay's positive charge sites virtually become part of it and are quite definitely 'blocked' by it as shown in Figure 1.13b. In this condition individual clay particles repel one another and the clay's contribution to soil structural stability is lost.

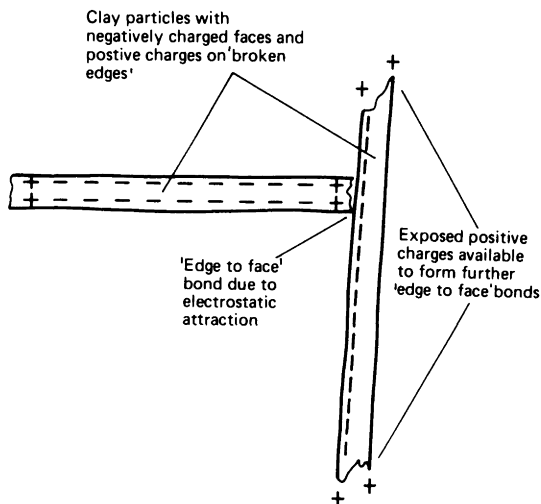


Figure 1.12 'Edge to face' bonding in flocculated clay

Calcium on an exchange complex is tightly bound, Ca^{2+} ions are to be found in the Stern layer rather than in the double layer. Consequently Ca^{2+} does not prevent flocculation in the way that sodium does (Ca^{2+} forms a compact double layer system, even with a dilute external solution). Thus the clay in a well limed soil can be relied upon to remain flocculated and therefore stable.

Prevention of deflocculation after salt water flooding

Deflocculation in salt water flooded soils is prevented by applying gypsum (CaSO_4) to ensure that the Na^+ is replaced by Ca^{2+} before the electrolyte concentration in the soil solution is allowed to fall.

Soil acidity and pH

The soil's acidity, measured on the pH scale, is of fundamental importance to farmers. When pH is too high, the availability of most trace elements to crops is reduced, whilst at low pH (acid conditions) crop yields are reduced drastically, due to excessive availability of aluminium and manganese.

In formal terms, the pH of a solution is defined as the negative logarithm of the activity of H^+ ions. Soil pH values range from about 4.0 (very acid) to around 8.0. The chemist's concept of pH 7.0 being neutrality is not appropriate to soils, and pH 6.5 is the value at which a soil normally exhibits neither acidic nor alkaline properties.

Measurement of soil pH

A sample of soil is shaken with 2.5 times its own volume of demineralised water and the pH of the suspension determined with an electronic meter which relies on the effect of H^+ ions on a thin glass membrane. Alternatively a coloured dye solution which changes colour according to solution pH may be used. There are a number of difficulties relating to measurement and interpretation of soil pH.

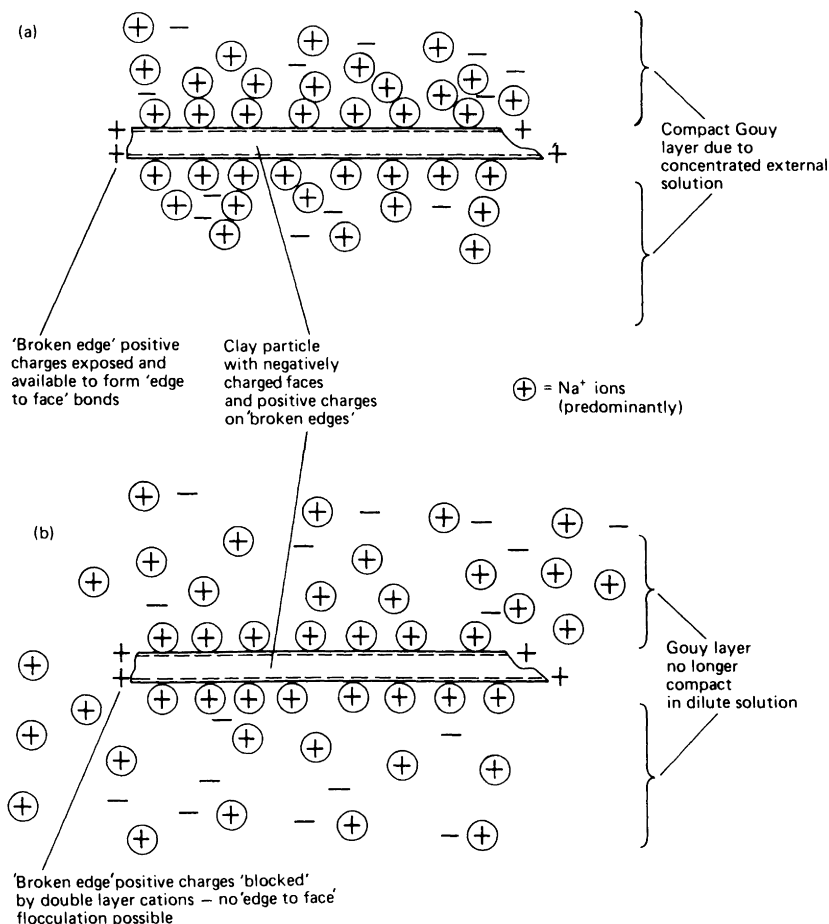


Figure 1.13(a) 'Broken edge' charges available for bond formation when the external solution has a high Na^+ concentration. (b) 'Broken edge' charges 'blocked' by the cations of the double layer with a weak Na^+ external solution

pH is particularly variable within fields and it is important to have a representative sample (see p. 28).

Duplicate *pH* measurements at the same point may vary over a time scale of days or weeks due to changes in moisture content, soil solution concentration or carbon dioxide level in the soil atmosphere.

The presence of a Gouy-Chapman double layer means that H^+ ion concentration is not uniform throughout the whole suspension.

Thus interpretation of a *pH* reading is imprecise — no one can know whether a particular root environment corresponds to the external solution or something near to the composition of the Stern layer.

Large variations in the soil:water ratio affect the result obtained. On average, the *pH* sensitive surface of a glass electrode is nearer to the inner (Stern) layer in a more concentrated suspension and so records a lower value (more H^+ ions) than in a more dilute suspension.

An electrical junction effect due to clay particles may occur in the reference electrode system of an electronic *pH* meter, giving a slightly erroneous reading. This effect is most

pronounced when measurements are made in thick suspensions or pastes.

Hydrogen ions are cations and therefore subject to the Gapon equation (see p. 14) so that their activity is controlled in part by the activity of other ions in the system. If, for example, K^+ ions are added to the solution, H^+ ions will transfer from the exchange sites into the solution, increasing its acidity (i.e. lowering the *pH*). When *pH* is measured by the water suspension method described, the cations in the external solution are derived entirely from the sample itself. The cation concentration will be governed by the previous history of the soil and the soil:water ratio. This uncertainty and its consequent effect on *pH* may be reduced by measuring *pH* in a salt extract having a concentration similar to that of a typical soil solution; 0.01 mol CaCl_2 is frequently used.

pH measurement using 0.01 mmol CaCl_2

Results obtained by using 0.01 mol CaCl_2 in place of water for making the suspension are generally about 0.5–1 *pH* unit lower at the acid end of the range. Differences are smaller at

high pHs. The Soil Survey of England and Wales routinely uses both methods and an idea of the relationship between them can be obtained from the analytical data on typical soils (see Table 1.17). Despite the theoretical advantage of CaCl_2 as an extractant for pH measurement, it is not used for agricultural advisory purposes in the UK — most people working with crops are familiar with the numerical values given by the 'water method', and the advantages of CaCl_2 are not thought sufficient to justify any change. Because of all the sources of possible error in measurement or interpretation, there are few practical circumstances where it would be advantageous to know a pH value to an accuracy better than ± 0.2 pH units.

The effects of soil properties on activity and degradation of pesticides

Adsorption

The concentration of a pesticide in soil solution is determined by the extent and strength with which it is adsorbed by soil colloids. This in turn determines its effectiveness or toxicity to weeds or insects, its rate of degradation, leachability and uptake.

Organic matter (OM) content is the soil property with the greatest influence on pesticide adsorption, so %OM must be taken into account when deciding the dose necessary to give the required activity for pest or weed control without causing crop damage. At one extreme, on soils of high OM level, the dose necessary to give the required effect may be prohibitively expensive or undesirable. For many pesticides adsorption by peat soils may be virtually complete, so the pesticide is entirely without effect. By contrast, on some sandy soils with low OM levels, certain soil-acting herbicides cannot be used because of the risk of damage to the crop.

The bipyridyl herbicides, paraquat and diquat, are strongly and irreversibly adsorbed by clay minerals and are thereby rendered inactive on contact with the soil. However adsorption onto organic surfaces is reversible, so that bipyridyl herbicides adsorbed by peats or humose layers can subsequently be taken up, and may damage crops.

Soil texture is closely related to both the soil's colloid surface activity and its ability to retain moisture. Consequently, texture is important in determining suitability and application rates for residual herbicides. Figure 1.14 illustrates this schematically.

Pesticide degradation

Most pesticides in common use are degraded by microbial action. Bacteria, actinomycetes and fungi are able to utilise even chemically stable pesticides as carbon sources. Typically, such microbial degradation is preceded by a 'lag phase', the time taken for a population capable of performing the particular degradation to build up in the soil. Thereafter, the soil is 'enriched' with adapted organisms which enjoy a competitive advantage over their non-adapted counterparts whilst the substrate (pesticide) is present. Subsequent additions of the same pesticide to enriched soil are degraded without any lag phase so long as the adapted organisms remain viable.

Some pesticides may be protected from microbial attack by adsorption onto soil organic colloids. Thus residue levels are frequently much greater in peaty soil than in mineral soils. Because these residues are adsorbed, they do not

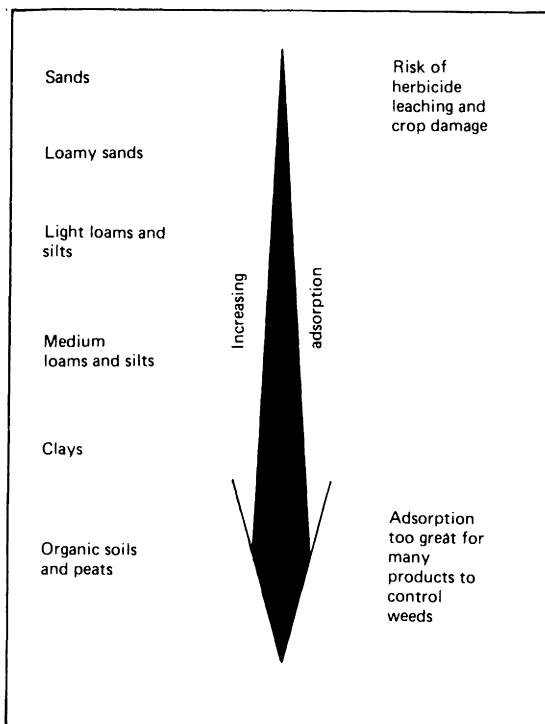


Figure 1.14 Effect of soil texture on adsorption of residual herbicides

normally support high concentrations in the soil solutions, so are not taken up by roots.

SOIL ORGANISMS

Chemical processes in the soil can be imagined as a number of cycles. Carbon is cycled between atmospheric carbon dioxide and plant tissue, and nitrogen is fixed from its gaseous form, incorporated in living tissue then released again to the atmosphere. Similarly, other elements are taken up by plants, become part of their structure and are eventually released to be utilised over and over again. All such cyclic processes in soil are dependent on the processes of decay and degradation brought about by living organisms.

Recently dead plant (or animal) remains are either consumed by animals, worms, slugs, etc. (the fauna) or decomposed *in situ* by smaller micro-organisms. Fungi generally mount the initial attack on resistant organic remains (which include the excreta of the faunal population), whilst bacteria and actinomycetes require chemically less complex nutrients such as simple sugars. Algae, at least when found on or near the soil surface, carry out photosynthesis in the same way as green plants, and some are able to fix nitrogen. The smallest of the fauna, the protozoa, feed largely by capture of other micro-organisms, particularly bacteria, whilst the larger macro-fauna (e.g. earthworms) feed on plant and animal remains. They also have a mechanical effect on the soil, improving aeration and drainage as well as mixing soil constituents.

Note that the various soil organisms are not to be regarded in isolation. All compete for food; some are predatory on others. Whilst some exist together in symbiotic relationships, others exude antibiotics which may confer some competitive advantage. The guts of saprophytic invertebrates commonly support a protozoal population to facilitate cellulose decomposition, and there are numerous examples of symbiosis between micro-organisms and the plant population.

Nutrition of soil micro-organisms

Historically micro-organisms have been thought of as either autotrophs, needing no organic substances, or heterotrophs needing organic compounds of various degrees of complexity. This classification leads to confusion, for example, when dealing with bacteria *capable* of functioning without any organic substances, but only showing maximum growth when obtaining energy from oxidation of carbon compounds. Such difficulties are avoided when it is understood that nutrients have three separate functions:

- (1) provide the building blocks from which the organism is formed;
- (2) provide a source of energy;
- (3) accept electrons released during the oxidation reaction which provides energy.

Autotrophic organisms are either chemoautotrophs, obtaining energy from inorganic oxidation reactions, or photoautotrophs carrying out a photosynthetic reaction similar to that of higher plants. As well as the complexity of their carbon source requirements, heterotrophs have differing electron acceptor requirements. In aerobes the electron acceptor is oxygen, whilst anaerobes utilise some organic products of their own metabolism or an inorganic substance.

Classification of soil micro-organisms

Classification may be according to taxonomy, morphology (form as revealed by observation) or physiology (function). In the case of soil bacteria the activity of the various species is well documented. Hence, from the viewpoint of soil studies, it is profitable to adopt a classification based on physiological criteria. When dealing with actinomycetes, fungi, algae and protozoa it is not always possible to ascribe clearly defined functions to the various species, and only taxonomic or morphological classifications are used.

Bacteria

Due to difficulties in reliably distinguishing bacteria from the rest of the soil, estimates of their population vary widely. The commonly accepted range is in the region of 10^7 – 10^9 organisms/g of soil, equivalent to a live weight of about 3 t/ha (i.e. less than 1% of the soil organic matter).

Morphological classification distinguishes between cocci (diameter about $0.5\ \mu\text{m}$ $\times$ $3\ \mu\text{m}$ length) and rods of diverse shape, but is of little use since many bacteria are able to change from one form to another and there is no correlation between morphology and function. A more useful type of classification for heterotrophs is that proposed by Lockhead

Table 1.3 Lockhead and Chase's classification of heterotrophic bacteria according to complexity of nutrient requirement

Group	Types included
1	Those growing in a simple glucose–nitrate medium containing mineral salts.
2	Those that can only grow when ten amino acids are added to this medium.
3	Those that can only grow when cysteine and seven growth factors (aneurin, biotin, etc.) are added.
4	Those needing both the amino acids and the growth factors.
5	Those needing yeast extract.
6	Those needing soil extract.
7	Those needing both yeast and soil extract.

and Chase (1943), based on nutrition requirements and summarised in *Table 1.3*.

It is observed that fertile soils contain a greater proportion of groups 5, 6 and 7 (having complex requirements), whereas the immediate root zone of plants is characterised by the presence of groups 2, 3 and 4. Large dressings of farmyard manure have been shown to increase populations of bacteria having the more complex requirements, whereas inorganic fertilisers are without any detectable effect.

The chemoautotrophic bacteria may be similarly classified on the basis of the oxidation reaction used for energy supply.

- | | |
|---|----------------------------------|
| (1) Nitrogen compounds oxidised | |
| (a) Ammonium oxidised to nitrite | <i>Nitrosomonas</i> |
| (b) Nitrite oxidised to nitrate | <i>Nitrobacter</i> |
| (2) Inorganic sulphur compounds converted to sulphate | <i>Thiobacillus</i> |
| (3) Ferrous iron converted to ferric | <i>Thiobacillus ferrooxidans</i> |
| (4) H_2 oxidised | Several genera |

The microbial oxidation of nitrogen compounds, in particular, is vital for crop nutrition *see p. 21*.

Bacterial fixation of nitrogen

Free living bacteria notably *Azotobacter* and *Beijerinckia* are capable of fixing atmospheric nitrogen, but quantities fixed are small. The process is inefficient in terms of nitrogen fixed per unit of carbohydrate oxidised and the rate of fixation appears to be limited by the carbohydrate supply. It is not therefore surprising that when such bacteria establish themselves in close association with the roots of plants which exude carbohydrate, they perform rather better. On maize, rice and some tropical crops, such associations can give fixation rates in excess 1 kg/(ha d) of nitrogen (N). Quantities of N fixed in this way under UK conditions are minimal.

The notion, once popular in Russian literature, that N fixation could be increased by inoculation with *Azobacter* has been shown to be fallacious — it is almost universally true that if the rate of a process is limited *only* by the size of a microbial population, that population will increase until some other factor, such as substrate availability, becomes limiting.

Symbiotic nitrogen fixation is a characteristic of legumes

and a small number of non-leguminous plants. Agriculturally, interest is confined to the legumes and in the UK the crops concerned are clovers, lucerne, peas and beans. The fixation occurs in root nodules formed from plant tissue, in response to invasion by *Rhizobium* bacteria. Up to 200 kg/ha of N may be made available to a UK clover crop in this way and rather lesser amounts to peas and beans. In grass-clover swards, grass is able to benefit from the 'clover nitrogen', although the mechanism involved has not been clearly established. In such mixed swards the *Rhizobium* can only start to fix N in the spring time once clover growth has become sufficiently vigorous to provide a carbohydrate supply.

Although attractive as a means of providing herbage without using high inputs of nitrogen, grass/clover swards have substantial management problems. The N supply from rhizobium does not become available until well into the growing season. Whilst use of fertiliser nitrogen can 'fill the gap' this may prove undesirable since N inhibits fixation and may also have a detrimental effect on clover survival. Clover is also very sensitive to grazing management and considerable skill is needed to maintain the clover content of a grazed sward. Current work at the Welsh Plant Breeding Station is addressing the joint problems of early growth and survival and leading to the development of early growing matched grass and clover varieties. Such developments should introduce new opportunities for exploitation of biological N fixation.

Bacteria and soil structure

Polysaccharide gums are thought to be of great importance in stabilising soil structure (see p. 10). Most of these are thought to be formed as products of bacterial metabolism.

Actinomycetes

In any strict sense actinomycetes are classified within the bacteria, but on the basis of morphology and physiology they represent a transition between true bacteria and the fungi. A large number of genera of actinomycetes have been recognised in the soil. Many of these are of interest because of their ability to synthesise antibiotics and the soil has been the primary source of organisms for commercial production of these agents in the pharmaceutical industry.

In agricultural soils, abundance is in the order 10^5 – 10^8 /g but only *Streptomyces* (70–90% of the population) and *Nocardia* (10–30%) are present in sufficient numbers to be of importance in soil studies. The nocardias are sufficiently similar to bacteria that they are frequently (and perhaps inadvertently) included in estimates of bacterial populations. Some of their number have the ability to modify their metabolism quite rapidly, to decompose synthetic chemicals.

Streptomyces are responsible for the characteristic smell of freshly turned earth. The role of actinomycetes in soil processes is still not clearly defined, but it appears that they have a lesser biochemical importance than bacteria and fungi. They have been observed to attack resistant components of plant and animal tissues some time after addition of fresh remains, presumably when competition from organisms with more fastidious nutritional requirements has declined. Their metabolites are of interest since they are very similar to the complex molecules described as 'humus' when occurring in the soil. Actinomycetes are favoured by abundance of organic materials, relatively dry

soil conditions and high soil pH. Under waterlogged conditions or at pHs less than 5.0 they are virtually absent and some authorities claim that it is possible to estimate the pH of arable soil on the basis of the smell associated with *Streptomyces*.

Thermoactinomyces and some species of *Streptomyces* have an important degradative function at elevated temperatures (50–65°C) in compost heaps.

Fungi

Fungi occur in all soils and they are commonly associated with the initial breakdown of organic debris. Many are saprophytes, tolerant of acidic conditions and capable of flourishing in the litter layer of forest soils. They are highly efficient at converting organic carbon into fungal tissue, even to the extent of taking nitrogen from the soil solution in order to metabolise organic matter with a large C:N ratio. Since fungal tissue is quite resistant to breakdown, N locked up in this manner may be unavailable for some time. Like the actinomycetes, fungi can incorporate N into humic-like substances which normally have a very long half-life in the soil.

Fungi are heterotrophs and require oxidisable carbon compounds as energy sources. The requirements of various species range from simple carbohydrates, sugars, organic acids and starch through cellulose and lignin (a substance resistant to attack by most other organisms) to those requiring complex chemical growth factors, or only capable of growing as parasites on living plant tissue.

The soil is a highly competitive environment and there are many organisms capable of metabolising simple sugars. Those of the fungi which have sugar as substrate maintain a competitive advantage by growing very rapidly, so consuming the sugar before it is lost to other organisms. The cellulose decomposers, Ascomycetes, Fungi Imperfecti and Basidiomycetes face less competition and have slower growth rates. Since there is virtually no competition for lignin as a substrate, the higher Basidiomycetes which function as lignin decomposers are able to grow very slowly and appear to build up a food reserve prior to attacking any new lignin source. Apparently the initiation of decomposition of lignified material is an energy-consuming process.

Common parasitic fungi cause several well known diseases of crop plants (see p. 264). Some other fungi are normally saprophytic, but may occasionally invade living tissue and produce disease symptoms. The mechanisms controlling this 'occasional parasitism' are not understood, but adequately fertilised, vigorously growing crops enjoying good cultural conditions rarely seem to be affected.

Symbiotic fungi — mycorrhiza

For many years it has been appreciated that a symbiotic relationship exists between some fungi and forest trees. The commercially important relationships such as occur between Scots pine and fungi such as *Boletus*, *Amanita* and *Lactarius* have been carefully studied. Generally the fungus is one requiring sugar as substrate and the most abundant supplies of sugars are found in trees whose protein synthesis is limited by nitrogen or phosphate deficiency. Thus the fungus invades the root tissue of the deficient trees and then 'repays' the tree by increasing the efficiency of uptake of N and P.

The relationship is in delicate balance: if the fungus is too successful supplying N + P to the tree its sugar supply will be reduced, whilst if the growth of the tree is checked for any reason, the fungus may become parasitic.

A number of crop plants can sustain populations of vesicular-arbuscular mycorrhiza in their root systems. Although these mycorrhiza are able to contribute to phosphate uptake at very low soil phosphate levels they do not appear to have any beneficial function under normal agricultural conditions.

Fungi and soil structure

Although fungi lack any significant capacity to produce 'gummy' substances, some have considerable power to stabilise soil structure by the mechanical effect of their mycelia binding soil particles together.

Algae

Algae are photoautotrophs using light to provide energy and absorbing atmospheric carbon dioxide. Thus their preferred habitat is the soil surface. Strictly, many soil algae are facultative photoautotrophs having the ability to metabolise some carbohydrates, and significant numbers are found in sub-surface habitats. Since algae only have any recognisable competitive advantage when exposed to light at the soil surface, the significance of these sub-surface populations is not understood. Algae form a relatively small part of the soil biomass, typically 10–300 kg/ha (but up to 1500 kg/ha where a visible surface bloom occurs). When growing at the surface, multiplication of algae is only limited by lack of light, moisture or inorganic nutrients and competition for other factors does not affect them. Thus an algal bloom on soil is indicative primarily of moist surface conditions and a lack of crop competition for light. Algae have no significance in UK agriculture, but the blue-green algae in particular, are important as nitrogen fixers in rice paddies. Algae and lichens (symbiotic alga + fungus combinations) are important as primary colonisers of bare rock and they provide the sole source of organic matter in embryonic soils.

Protozoa

The Protozoa are the smallest of the soil fauna, distinguishable from all the organisms previously described by having means of locomotion. They are very small, 10–80 µm in length, to enable them to move in soil water films. Nutrition may be photoautotrophic (e.g. *Euglena*) and some may exist as saprophytes. However, the major mode of nutrition is by predation on bacteria, small algae, yeasts, etc. The estimation of protozoal populations is extremely difficult, but estimates of low precision are acceptable since their sole claim to agricultural significance is their role in reducing the numbers of organisms that constitute their prey. 'Order of magnitude' calculations indicate that the weight of the protozoal population is about 1% of the weight of bacteria in the soil.

Other animal life

The fauna is capable of near infinite sub-division. For the present purpose it merely needs to be pointed out that

anything that eats, be it elephant or earthworm, is participating in the chain of events whereby vegetation is transformed to yield soil organic matter and plant nutrients. Those animals which have their being within the soil mass, rather than above it, fulfil the additional function of physically mixing the soil constituents. In this context, earthworms are worthy of particular mention. They ingest organic remains and excrete an intimate mixture of soil and fresh 'humus', whilst leaving behind them channels which provide easy access for plant root growth, and improve both aeration and drainage. Where earthworms thrive, they dominate the faunal population, exceeding the sum of all other fauna by up to five-fold, on a weight basis. The three million or so worms which may be found under one hectare of deciduous forest weigh about 2 tonnes. Where earthworms do not dominate the fauna, total populations are normally much smaller and the soil appears 'lacking in life' when examined in a profile section.

Manipulation of micro-organisms

Since micro-organisms occupy such a fundamental place in the cyclic processes involving nutrient elements in the soil, the idea of improving crop nutrition by influencing microbial activity is an attractive one. Although much has been written, particularly in Russian literature, about inoculation of soil with organisms such as the nitrogen fixing *Azotobacter*, research in UK (and elsewhere) has failed to demonstrate any benefits. This is thought to be due to inability of added organisms to compete with 'native' populations in what is a very competitive environment.

A notable area of success is the use of inoculated seed for legumes and the general rule seems to be that this technique is beneficial in environments lacking suitable indigenous rhizobia.

At least a part of the success of traditional crop rotations is attributable to the way in which rotations influence micro-organisms and a current research goal is to introduce a more direct element of 'biological control' into the control of root diseases. The introduction of organisms which produce growth promoting substances in the soil is another possibility, but such organisms would need to be able to compete with the indigenous population.

THE MAJOR NUTRIENT ELEMENTS

These are nitrogen (N), phosphorus (P), potassium (K), magnesium (Mg) and sulphur (S). They are present in soils and are taken up by plants in relatively large quantities. Calcium (Ca) is also a major element, but is dealt with separately (p. 26) because of its importance in relation to liming.

The major elements have a role in building the structure of plants, whereas the micronutrients (see p. 27) are important in enzyme systems and contribute to the plant's function rather than its structure.

Nutrient offtake and nutrient cycles

The quantities of phosphate and potash removed by normal yields of crops are given in *Table 1.4*.

Table 1.4 Phosphate and potash removal by normal yields of common crops

	<i>Fresh yield</i> (t/ha)	<i>P₂O₅</i> (kg/ha)	<i>K₂O</i> (kg/ha)
Cereal – straw burnt or ploughed in	7 (grain)	55	39
– straw removed	7 (grain)	63	79
Sugar beet – tops ploughed in	40 (roots)	32	84
– tops removed	40 (roots)	63	246
Potato – haulm not removed	40	40	232
Oilseed rape	3	48	33
Grass – silage	50	70	240
– hay	5	30	90
Kale	60	72	300
Swedes	40	28	96

ADAS data.

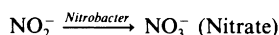
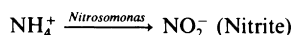
According to the type of farming, a part, or nearly all of the crop's major element uptake may eventually be returned to the soil in plant remains or animal excrement.

Nitrogen

Nitrogen exists in the soil either in the inorganic form, as ammonium (NH₄⁺) and nitrate (NO₃⁻) ions, or as organic compounds.

Inorganic nitrogen (also called mineral nitrogen)

This is the form of nitrogen used by crops. Nitrate (NO₃⁻) is the main form taken up by plants, and ammonium (NH₄⁺) is quickly converted in the soil to nitrate by the action of autotrophic bacteria:



These reactions can only occur under aerobic conditions. Since the second is potentially able to proceed faster than the first, NO₂⁻ does not normally accumulate in soils.

Nitrate ions are not held by soil colloids, so if not taken up by roots, nitrate is 'leached' by rainfall and lost in drainage water. Ammonium ions, however, are adsorbed by negatively charged exchange sites and so are held against leaching. Adsorption does not provide protection against oxidation to nitrate, so adsorbed ammonium is readily lost from the soil when conditions (temperature and aeration) are suitable for nitrification. Thus, with the exception of the driest parts of eastern England, it may be reliably assumed that between one autumn and the following spring, virtually all the organic nitrogen in a soil will be leached from the rooting zone unless taken up by a crop.

Levels of total mineral N (NH₄⁺ + NO₃⁻) in soils fluctuate widely over short time periods, so although it is a relatively simple matter to determine total mineral N in the laboratory, the result is of no use for determining fertiliser application rates. Levels normally fall in the range 5–50 mg/litre (mg of N/litre of soil), corresponding to about 10–100 kg/ha of N. The highest levels are frequently found under heavily fertilised grassland at the end of the growing season, before the onset of winter rains.

Organic nitrogen

Soil reserves of organic nitrogen are large. A soil with an OM level of 5% would have an organic N content of about 0.3%, or 6 tonnes/ha of N. This is some 60–600 times larger than the total mineral N.

The N in organic compounds is not available to plants until it has been released by the decomposition of organic matter. The fraction of the soil OM most actively involved in decomposition reactions contains most of its N in the form of proteins, and many micro-organisms are able to break these down to amino acids and then to ammonium ions, a process known as ammonification. In more general terms the conversion of nitrogen from organic to inorganic forms is referred to as mineralisation.

Since the small amount of mineral N is derived from the large heterogeneous reserves of organic N by biological degradative processes, it is not surprising that mineral N levels are highly variable. In general, soils with a high total N content are likely to need less N fertiliser to achieve maximum yields than those with low N levels, but the relationship is not good enough for predictive purposes. A 'total N' determination measures the reservoir of nitrogen potentially available to be mineralised, but tells nothing of the rate at which the micro-organisms are likely to carry out the mineralisation.

Nitrogen in crop residues and organic manures is not available to crops until it has been converted to ammonium. Subsequent conversion to nitrate is normally rapid and build-up of ammonium N only occurs under anaerobic (waterlogged) conditions. The ratio of C:N in the organic matter of agricultural soils in temperate lands is remarkably constant at between 10:1 and 12:1. If a manure or crop residue has a ratio less than this (i.e. more N relative to C) microbial action will release N quite rapidly. Materials with a large C:N ratio, such as cereal straw, require an additional source of N for their decomposition. In the absence of any other source, micro-organisms decomposing high C:N ratio materials will remove mineral N from the soil. Under present day conditions of high nitrogen fertiliser use this is not necessarily a bad thing. Post harvest mineral nitrogen levels in soil are usually high and this N is readily lost by leaching. However, when straw is incorporated this potentially wasted N may instead go to satisfy the micro-organisms' demands. This N is eventually released again when the biomass decays, so that as well as preventing pollution it can make a contribution to subsequent nutritional needs.