WORLD SOILS

THIRD EDITION

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CAMBRIDGE
UNIVERSITY PRESS

PUBLISHED BY THE PRESS SYNDICATE OF THE UNIVERSITY OF CAMBRIDGE
The Pitt Building, Trumpington Street, Cambridge CB2 1RP, United Kingdom

CAMBRIDGE UNIVERSITY PRESS
The Edinburgh Building, Cambridge CB2 2RU, United Kingdom
40 West 20th Street, New York, NY 10011-4211, USA
10 Stamford Road, Oakleigh, Melbourne 3166, Australia

First published 1970
Second edition 1978
Third edition 1997

Printed in the United Kingdom at the University Press, Cambridge

Typeset in Bembo 10/12pt

A catalogue record for this book is available from the British Library

ISBN 0 521 49777 9 paperback

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Preface to the third edition

The object of this book is to provide an introduction to the soils of the world and to give the widest possible readership a better understanding of their nature, properties and distribution. Environmentalists have vociferously drawn attention to the possible damage which can be caused by pollution of the atmosphere, the oceans, the rivers and aquifers, but all too often the effects of human actions on soils are overlooked. Very little publicity ever reaches the media about soils; they are not newsworthy and regrettably are regarded as unimportant by many people.

This is a pity because soils are some of the most interesting natural features on the Earth's surface. They are an essential part of natural ecosystems and are necessary for the growth of human food, animal fodder, fibre and timber crops. Soils underpin all natural and human-modified ecosystems to which they provide moisture, nutrients and support. It is just as important to have knowledge of and to care for our soils, as it is to study and be aware of the problems mankind faces if air and water supplies are abused.

Environmental awareness was raised by the 1992 United Nations Conference on Environment and Development (UNCED), even though those who prepared the documentation for this important conference scarcely mentioned soils; they did, however, refer to land and the concerns there are about the loss of agricultural production through land degradation, which is occurring throughout the world. Soil is a major component of land and, as such, the general public, administrators, planners and governments must be made aware of its importance. A significant outcome of UNCED was the idea of the sustainable use of the land. For soil scientists, soil conservation is not a new idea, but it must now be taken in a wider context embodied in the holistic approach introduced in Chapter 1. Whilst it is possible to dismiss this as simply a new way of looking at an old problem, soil scientists are keen to take the initiative in determining the most appropriate ways in which our soils are used.

Chapters 2, 3, and 4 follow the pattern of the previous edition of *World Soils*, describing the components of soil, the factors and processes which determine soil formation. Chapter 5, on soil classification, discusses some of the concepts behind classification systems and introduces the following seven chapters, which deal with the major soils of the world. Soils dominated by the nature of the parent materials are the subject of Chapter 6, and soils with moderately formed profiles are considered in Chapter 7. The increased importance of human modification of soils is reflected by an extensive review of Anthrosols in Chapter 8. The dominant soils of the cool, temperate, arid and humid tropical parts of the world are considered in Chapters 9, 10, 11 and 12 respectively.

The final two chapters are devoted to soil mapping and use of soil information. The skill and industry of soil scientists during the 20th century has produced soil maps of virtually all corners of the Earth and a wealth of knowledge and experience has been accumulated about the nature and properties of soils. Soil scientists wish to share this knowledge for the benefit of all human beings. Local, national and global environmental issues all require an input of basic soil knowledge and, with the encouragement of the International Society of Soil Science, and national soil science societies, this expertise is being made more widely available.

Education of future generations about the value of soils is particularly important, and consequently it is essential that the teaching profession is kept well-informed about soils. Soils can provide many interesting and different examples of investigative techniques for teachers who are involved with teaching basic science. Soil development and distribution have many intriguing aspects for geographical and environmental studies. Thus, the third edition of *World Soils* is a contribution to broadening the understanding of soils and raising public awareness of their importance. Many advances have been made during the past 15 years, so this edition has been completely rewritten to take into account the current understanding of our most basic resource.

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Acknowledgements

The author wishes to acknowledge the help and encouragement given by many friends and colleagues during the compilation of this edition. In the first place I wish to thank the Director and Staff of the International Soil Reference and Information Centre (ISRIC) at Wageningen for providing facilities for me to complete the revision of the text for this third edition. I am also indebted to ISRIC for the use of illustrative material from their worldwide collection of soils. A special word of thanks is reserved for my colleagues Dik Creutzberg and Hans van Baren who have both made constructive comments on draft chapters as they have been written. Their extensive knowledge of soils and wide experience have been of great help to me in presenting a picture of the soils of the world. I am also indebted to Otto Spaargaren, Sjef Kauffman, Albert Bos and Wouter Bomer for suggestions and help with the soil descriptions, analytical details and photographs of soils from their own collections and the ISRIC Soil Information System.

It is with great pleasure that I acknowledge the help of my former colleague Guy Lewis, of the Cartographic Unit at the University of Wales, Swansea, who has made rough sketches and vague ideas into the line diagrams which illustrate the text. The majority of the diagrams in this edition have been completely redrawn, but where close similarities exist to already published material, an acknowledgement is made in the figure caption.

A sincere word of thanks must be made to the many authors of books and scientific papers whose work has been consulted in the preparation of this edition; without their arduous field work and meticulous recording, a book such as this would not be possible. The inspiration of others is frequently responsible for triggering a chain reaction of ideas, and readers are strongly recommended to go back to the original sources whenever the subject interests them.

A short further reading list is presented at the end of each chapter, but reference has not been made to every factual detail used. As a very wide range of literature, together with considerable personal experience of soils on all five continents, has been used in preparing this edition, to fully reference each fact would make too many interruptions in the text and obstruct the presentation. As with the illustrative material, if a colleague finds a close similarity with some of his or her own concepts or ideas, I hope this general word of acknowledgement and thanks will be acceptable.

Lastly, and most importantly, I must acknowledge the support of my wife, who has endured many long absences whilst I have had the invaluable experience of working in different countries and gaining the background necessary for writing about the soils of the world.

1 Introduction

Soil is an important basic natural resource upon which we all depend. Farmers, horticulturalists and gardeners till it, providing human beings with food and beautiful flowering plants, foresters plant trees in it, engineers move it about in Juggernaut-like machines, small boys dig in it, and mothers abhor it as being dirty.

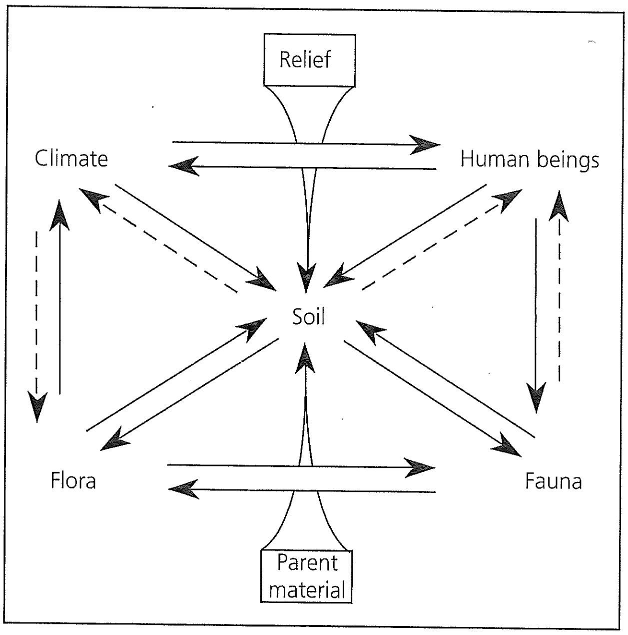
It is most unfortunate that soil, for many people, is simply trampled underfoot and is synonymous with dirt. They should know better, for soil plays a vital and important role in both the natural world and the life of human beings. It is one of the most undervalued of our natural resources, as virtually all human food is obtained from crops grown in soil, or from animals which graze on grass and other fodder crops also grown in the soil. Natural fibres such as cotton and flax, which provide us with clothing, are grown in the soil and soil materials can be used to make bricks, aluminium and glass. For many people throughout the world, charcoal and wood are the only sources of heat for cooking and comfort, and timber from trees is widely used for building purposes. Soils and soil products play an important part in our lives.

As Sir John Russell has written, 'a clod of earth seems at first sight to be the embodiment of the stillness of death'; however, he goes on to show that it is in fact a highly organized, physical, chemical and biological complex on which all of us are dependent. As the support for vegetable life, the soil plays the most fundamental role in providing sustenance for all animals and human beings, irrevocably linking the living world to the inanimate rocks and minerals.

The position of soil in the whole biotic complex can be illustrated diagrammatically (Fig. 1.1). It can be seen that the climate influences plants, animals and soils directly. However, plants influence the soil in which they grow, the animals which live in and on them as well as the climate near the ground. The animals, too, play a considerable role in soil development and often the nature of the soil in turn influences the animals which are present in it. Animals often exercise considerable control over the plants growing in the soil, and human beings may be responsible for the maintenance of the soil's fertility or its degradation. Finally, climate, through weathering, causes breakdown of the rocks, which in time become part of the soil as they are first weathered and later acted upon by the soil-forming processes.

The study of soils is of interest to many environmental scientists and in particular is the focus of study for the soil scientist. Soil science has several specialist branches, including the study of soil through physics, chemistry and biology. However, the aspect of soil science which brings all these points of view together is known as **pedology**. The pedologist studies soil as a natural

Fig. 1.1 The position of soil in the environment. The arrows indicate the various interactions and suggest the strength of the different influences.



phenomenon in its own right, and is interested in the appearance, the mode of formation, the distribution and the classification of soils wherever they occur.

In this respect, pedology makes use of a large number of branches of scientific knowledge, and as an integrative science resembles the role of physical geography. Chemistry, physics and biology have important contributions to make to the study of soils, as have studies in agriculture, forestry, history, geology, geography and archaeology. From all these subjects the information can be obtained for synthesis into a scientific discipline and natural philosophy separate from, and yet closely related to, many other branches of natural science.

Pedology may be studied as a pure science in which investigations into soil-forming processes, soil distribution and classification form an important part of the subject in their own right. However, the results obtained can be applied to practical problems in agriculture, horticulture, forestry, engineering, and more generally in planning the future use of the land. In this respect pedology, which draws from so many different disciplines, already has had to adopt a broad approach and so can claim to be both a pure and an applied science.

In all branches of science, one of the main reasons for investigation is to obtain information which can be applied for the benefit of mankind. Although in the past much interesting scientific work on soils has been accomplished, its results have not always been applied, nor have they provided the benefits which were anticipated. One reason for this is that soil is a much more complicated entity than can be dealt with by the scientific approach alone: there are also political, social, economic, financial and legal implications to be considered. It may be taken out of its natural context for study, but research into its problems and the advice given to land-users cannot take place in a scientific vacuum. These ideas have resulted in recent proposals to adopt a holistic approach for future studies in soil science. This approach recognizes that soils are used within a social and economic framework which cannot be ignored when framing scientific policy, or when providing advice to farmers, foresters or other users of the land.

The emergence of soil science

It is recorded that, about 4000 years ago, attempts were made in China to recognize different types of soils and to arrange them in a sort of land classification. In Europe, 2000 years ago, the Greeks and Romans followed a similar approach. The best-known Roman book on agriculture was that of Columella, in which good soils were distinguished from those which were poor for plant growth. This sort of practical knowledge was passed on by medieval scholars, who repeatedly copied what was known already without improving upon it. By the late 18th and early 19th centuries a scientific interest in the natural world had emerged from the alchemy of earlier years, and this eventually gave rise to the branches of science recognized today.

Present-day soil science emerged just over 100 years ago from two different schools of thought, one chemical, the other geological. The German scientist Liebig (1803–1873) was probably the most renowned of the early exponents of the chemical view of the soil, but even before Liebig, in 1803, a Swedish scientist, Berzelius, described soil as a chemical laboratory of nature in whose bosom various chemical decomposition and synthesis reactions took place in a hidden manner.

Early pedologists with a geological background considered soil to be comminuted rock with a certain amount of organic matter added from the decomposition of plants. As late as 1917, a German scientist, Ramann (1851–1926), described soil as 'rocks that have been reduced to small fragments and have been more or less changed chemically, together with the remains of plants or animals that live in it or on it'. Even today it is still possible to hear people use terms of convenience such as 'limestone' soils or 'granite' soils, but the soils which overlie these parent materials are very different from the rocks in almost all respects.

The origin of current definitions of soil results mainly from the work of two men: in Russia, Dokuchaev, and in the United States, Hilgard. In the 1880s, it was Dokuchaev (1846–1903) who first enunciated the concept of the soil as an independent natural body, but separately both men noted that soils could be described in broad geographical zones which, with the state of knowledge at the time and at the scale of world maps, could be correlated with climate zones and also with the associated belts of natural vegetation. Although we now know this to be only partly true, it did serve to direct attention to the environmental relationships of the soil cover of our planet.

In the USA, Hilgard (1832–1916) noted that soils in the state of Mississippi were distributed in broad belts associated with surface geology, but when he subsequently moved to work in California, he realized that climate also had an important influence, for instance on the accumulations of carbonates and other salts present in the soils of drier regions.

Definition of soil

Soil may be defined broadly as *the unconsolidated mineral or organic material at the surface of the Earth capable of supporting plant growth.* This is in line with a simple definition used in an agricultural context, which describes soil as *the stuff in which plants grow.* Such definitions are sufficient for many practical purposes, but for the soil scientist a definition of a more explanatory nature is preferable.

The definition propounded by Joffe (1949) has the advantage that it brings together the physical character, as well as the chemical and biological constituents, and throws the right amount of weight on the morphology of the soil: *The soil is a natural body of animal, mineral and organic constituents differentiated into horizons of variable depth which differ from the material below in morphology, physical make-up, chemical properties and composition and biological characteristics.*

Another definition describes soils as *the collection of natural bodies, formed on the Earth's surface containing living matter and supporting or capable of supporting plants.* The action of various chemical, physical and biological processes on the original geological materials produces features which we recognize as soils. These changes, which affect the upper part of the weathered rocks (or regolith), can be seen as a sequence of layers, technically called **horizons**, having different colours, composition and structure in a sequence known as the **soil profile**.

The soil profile

A soil profile comprises the natural horizons of the soil as revealed in a vertical section, from the organic material at the surface, through the horizons of the mineral soil, to the parent material or other layers beneath, which influence the genesis or behaviour of the soil (Fig. 1.2).

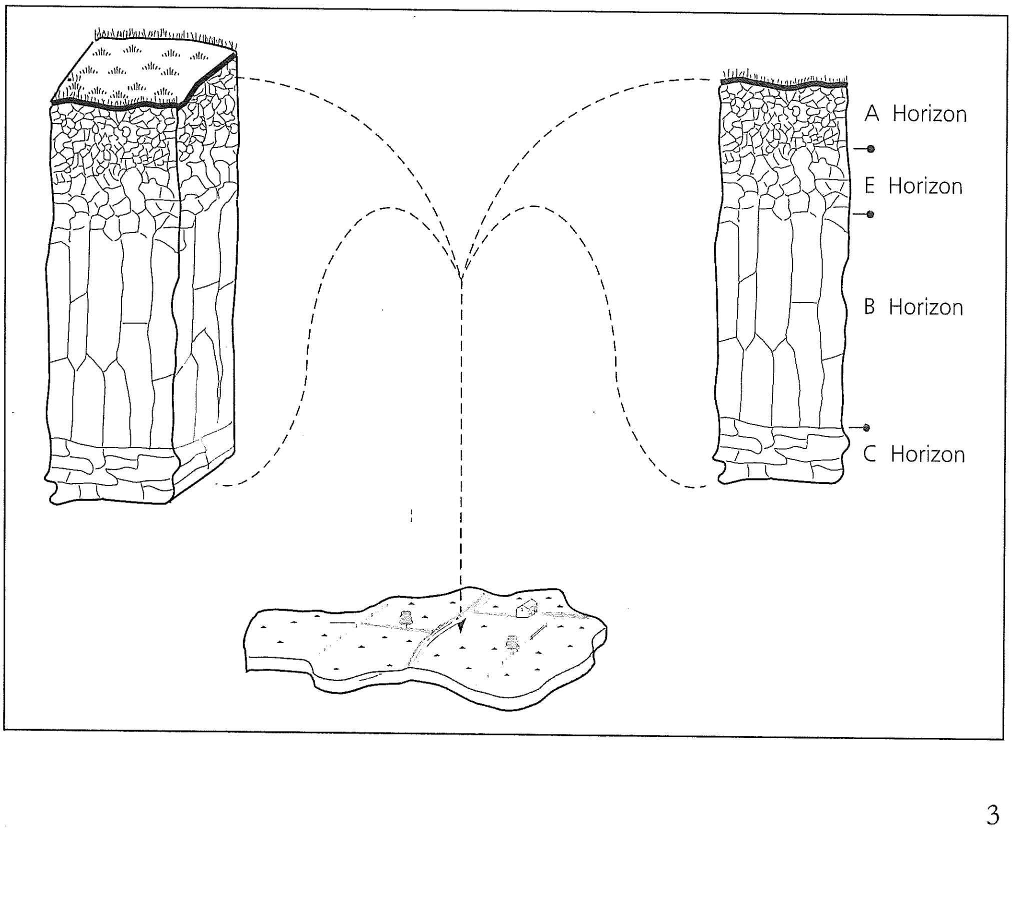


Fig. 1.2 An area of land with its soil shown as a pedon (left) and as a profile (right).

This concept of the soil profile is a simple view of the soil in a vertical direction, whereas soils are in fact three-dimensional features, extending over the landscape. Thus, for the purpose of classification and mapping, pedologists have adopted the idea of a soil volume, a three-dimensional body of soil called a **pedon**. A pedon is described as the minimal amount of material which can be logically called 'a soil', and it may range in size from 1 to 10 m^2. As soil is rather variable in composition and appearance, a range in characteristics is encompassed by allowing a number of pedons to be grouped together as a **polypedon**, which is a recognizable, geographic area of closely related soils. For practical purposes, most pedons described in the literature are of the soil revealed in a profile pit, the lateral dimensions of which are approximately 1 by 1.5 m, and its depth is governed by the depth of the soil. It is only in extended sections, such as road cuts or pipeline trenches, that the true picture of lateral variability of soils can be appreciated.

Soil horizons and their designations

A soil horizon is a layer of soil, revealed in a soil profile, lying approximately parallel to the Earth's surface, which possesses pedological characteristics.

The horizons of a soil profile are the morphological expression of the processes which have formed the soil. These processes will be discussed in Chapter 3. Horizons can be distinguished one from another by their physical make-up, chemical properties and composition, and biological characteristics. The vertical and horizontal limits of horizons occur where these attributes are significantly different in appearance or amount.

Certain soil horizons possess specific characteristics resulting from their mode of formation which, through experience, pedologists think are significant. These horizons can usually be readily identified in the field and they bring a sense of order to the description of the soil profile and permit the identification of the horizons present.

Early in the development of pedology, Dokuchaev began a system of labelling the horizons which he observed in the soil profiles he was describing. At first he used either letters (ABC) or numbers (123) as ciphers simply to label the horizons, but by the beginning of the 20th century these letters had acquired a special genetic significance. These horizon designations are still in use today. Over the years their purpose has been questioned and their role redefined, but whatever their detractors say, horizon designations are of great use in soil description, interpretation, in discussion and even in classification of soils.

Two systems of horizon designation, which are very similar and in wide usage, are those of the Food and Agriculture Organization of the United Nations (FAO) and the United States Department of Agriculture (USDA). As this book is concerned with a review of world soils, the FAO system will be presented here. Seven capital letters, O, H, A, E, B, C, and R, are used to represent the most significant horizons, called **master horizons**. Subordinate characteristics within the master horizons are designated by a lower-case letter following the master horizon letter. Each country with a soil survey organization may use these letters in a slightly different way, and give the letters their own particular meaning.

A summary of the master horizons and the main subdivisions of the FAO system is given in Table 1.1.

The O and H horizons are often replaced by the letters **L** for litter, **F** for fermentation and **H** for humus horizons, but the other master horizon designations are now in standard use in virtually all parts of the world. These designations will be used throughout this book and will be shown by the symbols depicted in Fig. 1.3.

The changing paradigm of soil

Unlike an animal or a plant, a soil is not a separate organism. Its lateral boundaries to other soils are not usually abrupt, but zones of transition. Although it is possible to see without difficulty the upper surface of a soil, the lower boundary between soil and 'non-soil' material is rather vaguely defined. In the preceding discussion of the soil profile, it was stated that the simple section or soil profile was an inadequate way of looking at the soil and that pedologists had adopted a three-dimensional approach. A well-known American pedologist has written that 'the success of a teacher

O An organic horizon at the soil surface, normally not saturated with water.

H An organic horizon at the soil surface normally saturated with water, characteristic of peaty deposits.

A A mineral horizon formed at or near the surface, characterized by the incorporation of humified organic matter intimately associated with mineral materials. Subdivisions include:

- **Ah** for an uncultivated horizon; accumulation of humus;
- **Ap** for a cultivated (ploughed) horizon;
- **Ag** for a poorly drained surface horizon.

E A mineral horizon, just below the soil surface, which has lost clay, organic matter or iron by downward movement. Subdivisions include:

- **Eg** for poorly drained horizons.

B A subsurface mineral horizon resulting from the change *in situ* of soil material or the washing in of material from overlying horizons. Subdivisions include:

- **Bg** for poorly drained;
- **Bh** for accumulation of humus;
- **Bs** for an illuvial accumulation of iron or aluminium sesquioxides;
- **Bt** for increase of clay;
- **Bw** for changes of colour or structure;
- **Bx** for compact brittle horizon known as a fragipan;
- **By** for accumulation of gypsum;
- **Bz** for accumulation of salts more soluble than gypsum.

C An unconsolidated or weakly consolidated mineral horizon which retains evidence of rock structure and lacks the properties diagnostic of the overlying A, E or B horizons. Subdivisions include:

- **Cg** for poorly drained;
- **Ck** for enrichment with calcium carbonate;
- **Cm** for cemented material;
- **Cx** for compact brittle material known as a fragipan;
- **Cy** for enrichment with gypsum;
- **Cz** for accumulation of salts more soluble than gypsum.

R Continuous hard or very hard bedrock.

Table 1.1 *Soil horizon designations*

depends to a high degree upon his ability to create in the minds of his students an integrated model of his subject as a whole'. Throughout this book a conscious attempt is made to provide the reader with such a model of the different major soils of the world and how they are related to each other.

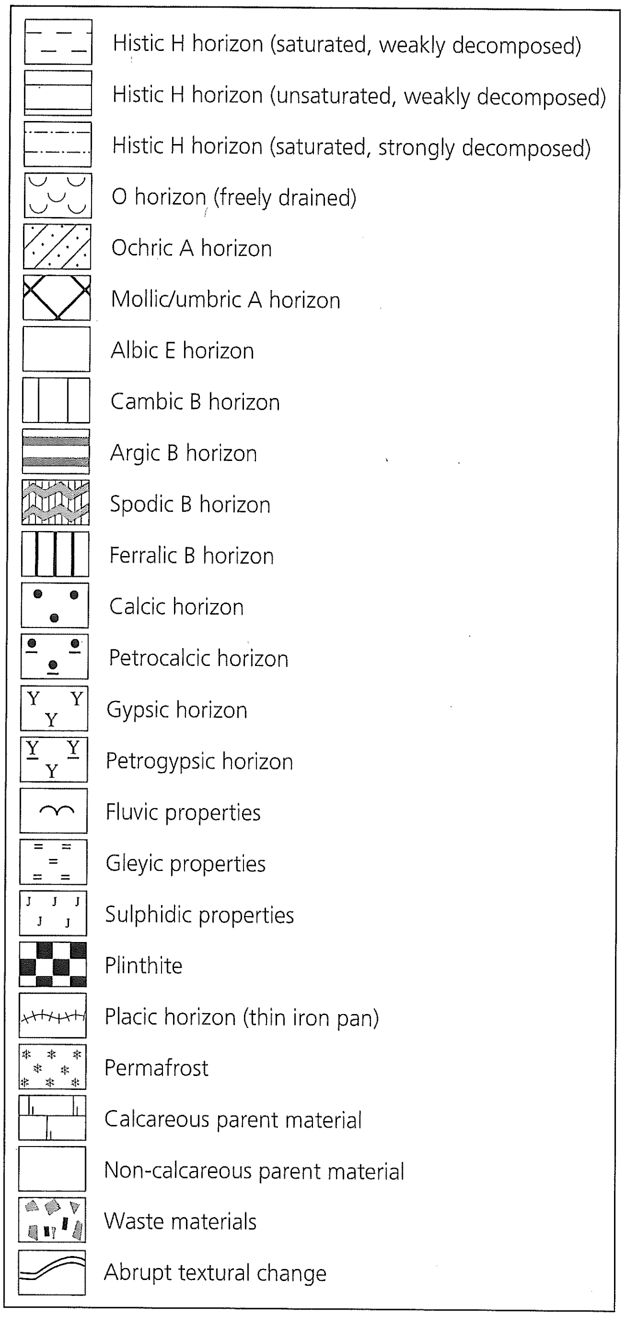


Fig. 1.3 Symbols used to represent soil horizons throughout this book.

From their studies of soils on the Russian plain, the Russian pioneers of modern soil science adopted the soil profile as the first 'model' of soil. As the effect of relief in Russia played only a small role in soil formation, the two-dimensional model sufficed, with the horizons developing parallel to the soil surface and only a small amount of lateral variation of soil characteristics. It was from an idea originally put forward by Dokuchaev that the first real conceptual model of soil formation was proposed by Jenny and published in 1941 in his book *Factors of Soil Formation*. This model, which will be discussed more fully in Chapter 3, was the most widely used model of soil development for almost 40 years. Another important general 'model' for soil development, proposed by Simonson in 1959, allowed a systems approach, with feedback for certain stabilizing functions which help to maintain the status quo. Many other models developed in recent years have permitted the estimation of certain unknown parameters from known measurements.

The widespread adoption of electronic methods of data storage in the last two decades has revealed shortcomings inherent in previous methods of soil survey, which gathered data for presentation in book and map format. These were usually a mixture of qualitative and quantitative data, providing information on the attributes of typical examples of soils from an area; they did not provide information on the continuous variability within soil mapping units. The nature of the data collected, in terms of classification units with abrupt boundaries, is not very suitable for input into data files for use in geographical information systems. It is argued by some authors that forms of continuous classification systems using 'fuzzy' mathematical methods should be used instead, but the fact remains that most available data were not collected for input into databases or computer models, and new data are either not available, or are very expensive to obtain.

Already, computerization of data has introduced a greater degree of standardization in the gathering of soil information worldwide. In recent times, many different models have been suggested which simulate processes operating in soils. The lead has been taken by studies of the influence of the physical characteristics of soils on the movement of water, plant nutrients and pollutants through the soil, their availability and effects on plants. However, extrapolation from these detailed site-specific data to regional or global systems can be fraught with many difficulties.

It will be obvious that soil is regarded in different ways by different people; attention was drawn to this in the opening paragraph of this chapter. Soil scientists are no exception to the public at large, and individuals hold different points of view. Views of the soil have also changed with time as new ideas have emerged and new facts have been discovered. Changes are constantly taking place in the way soil scientists regard the soil. Stimulated by the challenges provided by the United Nations Conference on Environment and Development (UNCED), held in Rio de Janeiro in 1992, ideas of soil resilience to some uses and the vulnerability to damage by other land-use practices have led to concepts of sustainable use of the land. In order that successive generations can have the means to grow crops, it is necessary to ensure that our current use of the soil will not detract from the future use of the land.

The realization that widespread damage has already occurred to the soils of the world, and that the warnings given by responsible soil scientists, amongst others, have not been heeded, has led to the suggestion that a wider, more comprehensive approach to soil conservation should be adopted, involving the users of soil in a closer understanding and collaboration with other scientists. This **holistic** approach may be defined as *the task of all people concerned with the soil to direct their interest, not just towards the physical, chemical and biological aspects, but also to those environmental economic, social, legal and technical aspects that affect soil use.* In the past, the results of research in soil science have not been effectively communicated to those who work the land, those who plan its future or those who administer the legal and financial framework within which soils must be conserved. This situation must be changed.

Land, with its cover of soil, has always been an important element in the environment of the human race; it has supplied food, fibre and forestry products. These crops also provide a renewable source of energy. The soil is a biological factory for waste purification; it has a rich diversity of flora and fauna which can convert most organic compounds into simple recyclable elements; the same flora and fauna help soil to maintain clean, fresh water supplies both on the surface and in the deep aquifers. Soils, through their carbon content, can influence the atmospheric concentration of 'greenhouse gases' such as carbon dioxide, methane and nitrous oxide. Soil is also the surface upon which almost all human activity takes place. Soil is clearly one of the most important components of the terrestrial environment.

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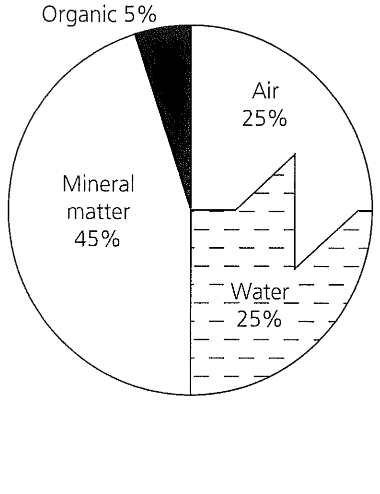
2 Composition of soils

There are four main constituents of soil: mineral matter, organic matter, air and water (Fig. 2.1). These will be considered in turn in this chapter. The mineral matter includes two main groups of materials. First, there are the resistant minerals weathered from the rocks and which persist to form the parent material; these are often referred to as primary minerals. Secondly, there are those minerals which have been formed in the soil by recombination from substances in the soil solution. The former are often coarse-grained, the latter fine-grained. The presence of sand, silt and clay particles in different proportions determines the texture of the soil.

Organic matter is derived mostly from decaying vegetable matter which is broken down and decomposed by the action of the many different life forms which inhabit the soil. Composed mainly of cellulose, starch and lignin in various states of decomposition, the organic matter of soils takes on several distinctive forms.

Except for loose sands and a few other soils which are described as structureless, the mineral and organic components of soil are aggregated in discrete structural units, termed peds, which are surrounded by spaces. These spaces between the peds, and spaces between individual sand grains and organic soil constituents, are referred to collectively as pores. Normally, both air and water occupy these spaces between the mineral aggregates, but if the soil is saturated most of the air is driven out. In a soil which is freely drained, some water is still present as thin films around the mineral particles, leaving the pores open for gases to diffuse in and out of the soil.

Fig. 2.1 Volume composition of a typical topsoil: amounts are approximate as the percentage of certain constituents (e.g. air and water) is constantly changing.



Mineral matter

The mineral portion of the soil is derived mainly from the geological substratum by processes of *in situ* weathering. However, in some locations, the mineral material from which the soil is formed has been transported by a river and laid down as alluvium, or it has moved downslope by creep as colluvium. In addition, a variable contribution comes from fine aeolian dust. The mineral part of soils consists of particles with a range of sizes, from very small clay particles up to sand-size particles. This part of the soil is known as the **fine earth** and it is upon this fraction of the mineral material that the **texture** of the soil is based. Particles larger than 2 mm, stones or gravel, also occur, but except for their physical presence and bulk they contribute little to the soil.

The fine earth, which most people think of as soil, is composed of three fractions, sand, silt and clay, determined according to the size of the particles. In an internationally agreed logarithmic scale, sand particles have a diameter of between 0.02 and 2.0 mm, silt has a size range of between 0.02 and 0.002 mm, and clay is the material of less than 0.002 mm in diameter. By taking a soil sample, moistening it and then estimating the proportions of sand, silt and clay present, it is possible to relate the soil to a texture class, as shown on the triangular diagram (Fig. 2.2). Descriptions of the twelve different classes of soil texture in common use are given in Table 2.1.

The particle-size classes are used for the description of texture of each horizon within a profile, but sometimes there is a need for broader groupings of soil texture for use when describing the whole profile or groups of profiles of similar soils. These textural groupings are:

Clayey	very fine
	fine
Silty	fine
	coarse
Loamy	fine
	coarse
Sandy	

The relationship of the twelve texture classes to these broader groupings can be seen in the triangular diagram (Fig. 2.2b). These broad groupings cut across several of the textural classes as they are based on slightly different particle-size ranges. This has occurred because of the need for closer co-operation between engineers and soil scientists, which has resulted in the pedological size grades being modified.

In turn this needed some modification of the texture classes to conform with different size grades. The scheme adopted by the Soil Survey of England and Wales is given in the third triangular diagram (Fig. 2.2c). Where stony material amounts to more than 35 per cent of the volume of a soil horizon it is indicated by the term **skeletal** (where rock fragments exceed 90 per cent the word **fragmental** is used by American pedologists) in association with the appropriate textural grouping.

These different soil textures reflect properties which influence the management and economic use of the soil such as permeability, water-holding capacity and aeration. Coarse-textured, sandy soils are usually freely drained, and in a dry summer may suffer drought, but cultivation is relatively easy. Frequently, clay soils are poorly drained, and the expense of installing a drainage system can be large. Clay soils are more retentive of plant nutrients, but cultivations are always likely to be difficult, although in a dry year these soils may produce

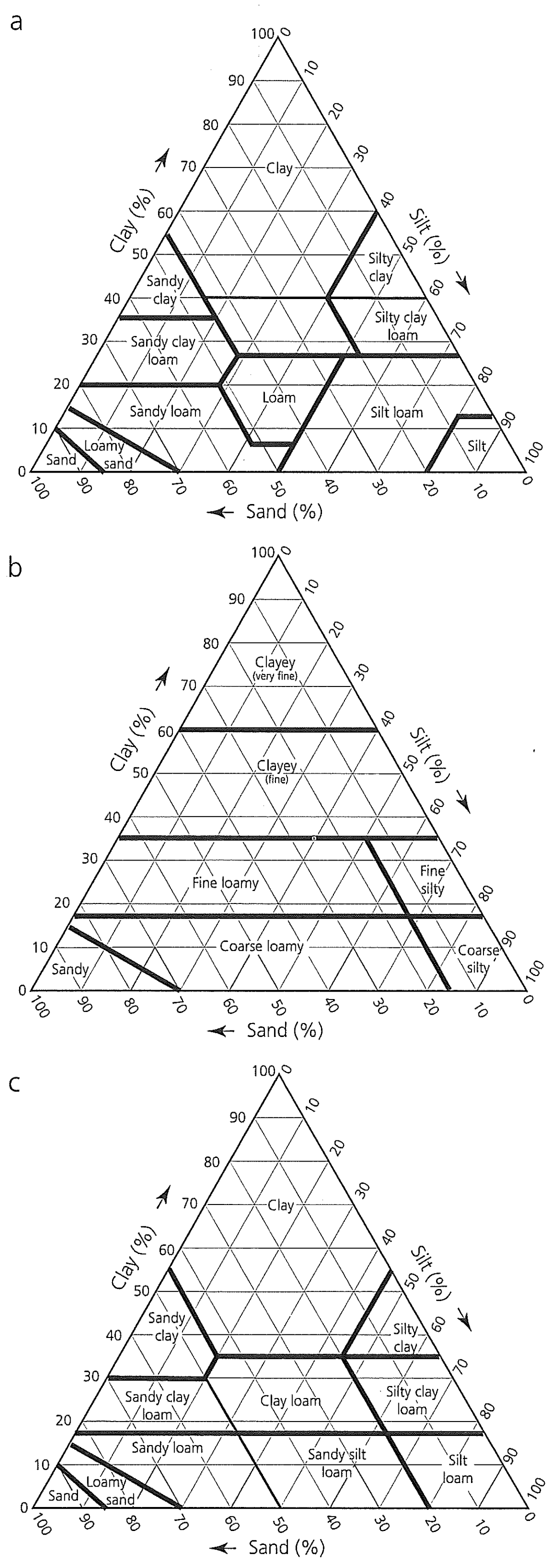


Fig. 2.2 Soil texture classes. The three sides of the triangle are base lines for sand, silt and clay with the opposite apex representing 100%. (a) USDA; (b) broad texture groupings, USDA; (c) soil texture classes, Soil Survey of England and Wales.

Sand. Soil consisting mostly of coarse and fine sand, and containing so little silt and clay that it is loose when dry and not sticky when wet. When rubbed it leaves no film on the fingers.

Loamy sand. Soil consisting mostly of sand but with sufficient clay and silt to give slight plasticity and cohesion when very moist. Leaves a slight film of fine material on the fingers when rubbed.

Sandy loam. Soil in which the sand fraction is still obvious, which can be moulded readily when sufficiently moist, but in most cases does not stick appreciably to the fingers. Threads do not form easily.

Loam. Soil in which the fractions are so blended that it moulds readily when sufficiently moist, and sticks to the fingers to some extent. It can with difficulty be moulded into threads but will not bend into a small ring.

Silt loam. Soil that is moderately plastic without being very sticky and in which the smooth soapy feel of the silt is dominant.

Sandy clay loam. Soil containing sufficient clay to be distinctly sticky when moist, but in which the sand fraction is still an obvious feature.

Clay loam. The soil is distinctly sticky when sufficiently moist and the presence of sand fractions can only be detected with care.

Silty clay loam. Soil that contains very small amounts of sand, but sufficient silt to confer something of a soapy feel. It is less sticky than silty clay or clay loam.

Silt. Soil in which the smooth soapy feel of silt is dominant.

Sandy clay. The soil is plastic and sticky when moistened sufficiently but the sand fraction is still an obvious feature. Clay and sand are dominant and the intermediate grades of silt and fine sand are less apparent.

Silty clay. Soil which is composed almost entirely of very fine material but in which the smooth soapy feel of the silt modifies to some extent the stickiness of the clay.

Clay. The soil is plastic and sticky when moistened sufficiently and gives a polished surface on rubbing. When moist, the soil can be rolled into threads, is capable of being moulded into any shape and takes clear fingerprints.

Table 2.1 *Soil texture class descriptions*

better crops than a coarser-textured soil. Silty soils may also be troublesome as they must be cultivated within certain moisture limits, otherwise they become cloddy and the preparation of a seed bed is made difficult. Also, the effect of heavy rain on silts can cause surface sealing which inhibits seedling emergence and encourages erosion.

Soil mineralogy

The minerals present in the soil belong to two completely different groups, based on their origin. **Primary minerals** are those derived from the parent material and **secondary minerals** are the product of chemical weathering and have been formed in the soil itself. The primary minerals present in a soil have usually been through at least one cycle of weathering, so that only the more resistant ones remain. In the soils of the humid temperate climate zone, the sand fraction is composed largely of quartz grains, but it may also contain some grains of feldspar, mica, and a few of the rarer persistent minerals such as zircon, tourmaline or glauconite (Fig. 2.3). These minerals can sometimes be used to determine the origin of the soil parent material and also to determine quantitatively the path of weathering from rock to soil. Quartz grains often constitute between 90 and 95 per cent of all the sand and silt particles of soils derived from sedimentary rocks.

In soils developed in the humid tropical environment, most weatherable primary minerals such as feldspars and mica are normally absent and the soil is composed largely of quartz and secondary clay minerals. The clay mineral kaolinite together with iron and aluminium hydroxides constitute the clay fraction. The amount of quartz sand is variable but silt-sized material is proportionately low.

The **clay minerals** are the most important of the mineral constituents of soils; they consist of fine ($<2\,\mu m$), platy-shaped mineral grains which can be identified only indirectly, or by an electron microscope. Clay minerals are characterized by a layered, crystalline structure, and chemically they are hydrous silicates of aluminium. There are three main members of this group of minerals: kaolinite, smectite and the hydrous micas, although transitional forms occur between each group. All clay

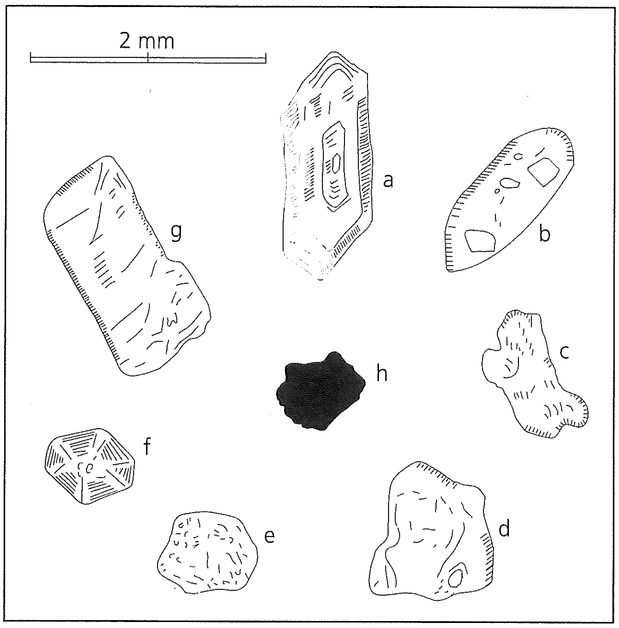
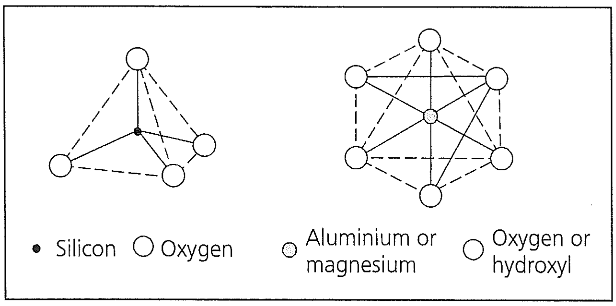


Fig. 2.3 Some of the less common (heavy) minerals present in the soil: (a,b) zircon, (c) glauconite, (d,e) garnet, (f,g) tourmaline, (h) magnetite.

minerals are constructed from layers of silica and aluminium atoms and their attendant oxygen and hydroxyl groups (Fig. 2.4). The silicon and aluminium layers are held together by shared chemical bonds. Kaolinite has one layer of silicon atoms and one layer of aluminium atoms in a 1:1 structure (Fig. 2.5). Smectite minerals have three layers with the aluminium atoms lying between two layers of silicon atoms in a 2:1 structure, sharing the valencies of their oxygen atoms.

Many soil minerals are derived directly from the parent material, but clay minerals are formed in the soil itself; these are referred to as secondary minerals. Weathering releases elements that pass into the soil solution from which recrystallization takes

Fig. 2.4 Silicon tetrahedron and aluminium octahedron: the building blocks for silicate clay minerals (after Brady, 1984).



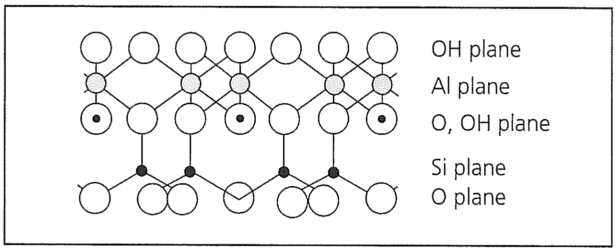


Fig. 2.5 The structure of kaolinite: sheets of silicon tetrahedra and aluminium octahedra are linked by shared oxygen atoms. It is a 1:1 mineral (after Brady, 1984).

place to form a completely new mineral. Hence kaolinite, $Al_4Si_4O_{10}(OH)_8$, can be formed in this way from soil solutions rich in aluminium and silicon; where base-rich conditions prevail the mineral montmorillonite, $NaMgAl_5Si_{12}O_{10}(OH)_6$ (smectite group), may be formed (Fig. 2.6). Clay minerals can also be formed in the soil by simple alteration of primary minerals.

Other secondary minerals can accumulate in the soil. Where oxidation and reduction are prevalent, iron and manganese oxides precipitate as scattered concentrations throughout the horizons of poorly drained soils as the minerals goethite or lepidocrocite. These concentrations grow by the addition of concentric layers of iron and manganese compounds. Where warm summer temperatures occur, the iron minerals may dehydrate to bright red hematite. Concretions of calcium carbonate or gypsum are unusual in the soils of humid regions because leaching prevents their formation, but they are common in the soils of seasonal and dry climates, such as on the steppes or in desert and semi-desert areas.

Because the clay mineral particles are so small in size, the minute electrical forces of the molecules at the surface of the clay become dominant and confer upon the clay particles a **colloidal state**. A colloidal state occurs when particles of less than one micron (<0.001 mm) are dispersed evenly throughout another medium. (Two familiar examples of colloids are milk, in which tiny solid particles are dispersed throughout a liquid, and cloud, where water droplets are dispersed in air.) The properties which the colloidal state confers upon soil materials are plasticity, cohesion, shrinkage, swelling, flocculation and dispersion, all of which are significant for the physical properties of a soil (Fig. 2.7).

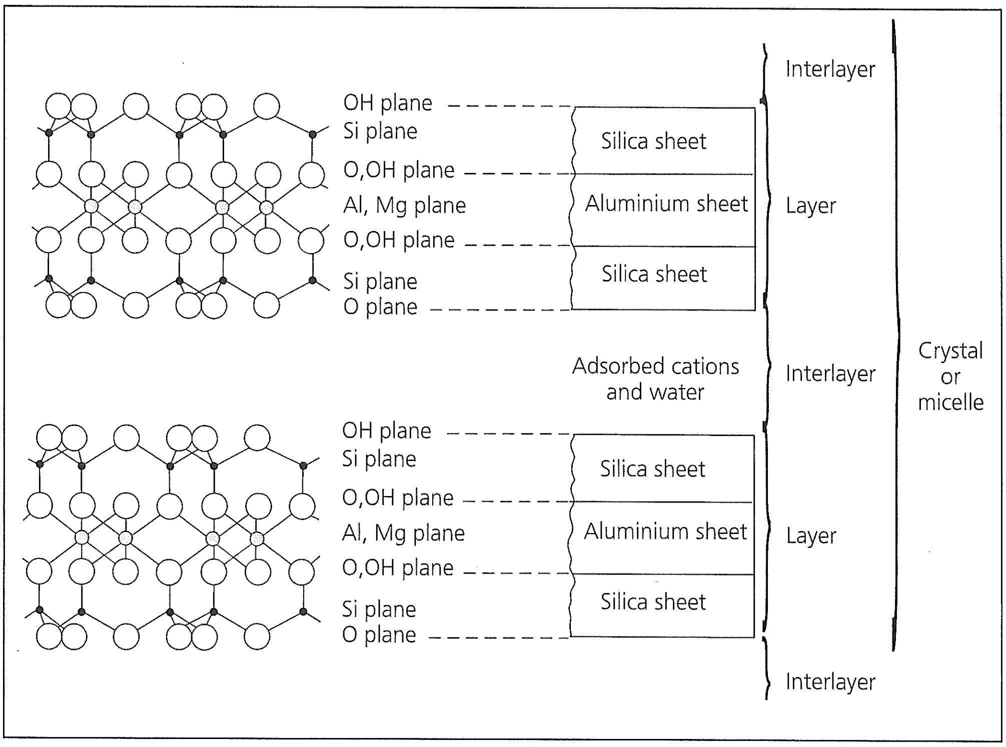


Fig. 2.6 Structure of montmorillonite (smectite): it is built of two sheets of silicon tetrahedra and one sheet of aluminium octahedra, linked by shared oxygen atoms. It is a 2:1 mineral (after Brady, 1984).

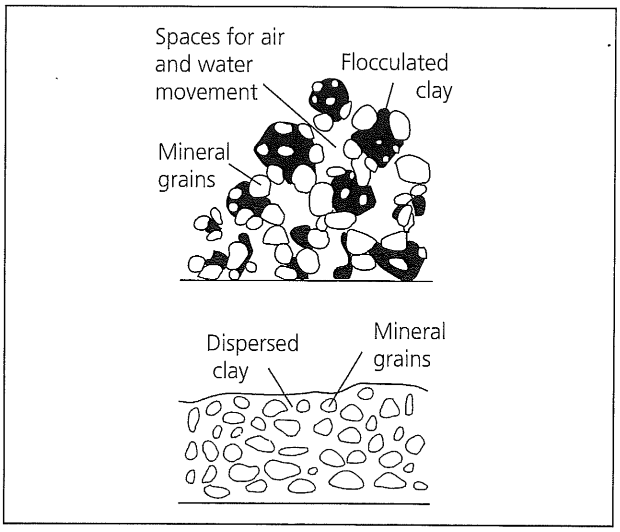
Clay minerals (and organic matter) have electrical charges associated with their surfaces. These charges enable clays to attract, hold and exchange cations and anions. The most important property of the clay minerals is the **cation exchange capacity** (CEC), measured in centimoles of positive charge per kilogram of soil. Kaolinite has a low CEC, 8 $cmol_c\,kg^{-1}$, but clays of the smectite group are higher, up to 100 $cmol_c\,kg^{-1}$; other clay minerals have an intermediate CEC.

Organic matter

Soil organic matter can be present in several forms: it may be intimately mixed with the mineral matter of the surface horizon, it may be present in an illuvial B horizon, or it may form discrete layers upon the surface. The four basic types of surface organic matter identified by pedologists are known as peat, mor, moder and mull. Gradational forms can be recognized between these four basic types. Peat is developed in saturated conditions, while the other three types of surface organic matter are only saturated for short periods of time (Fig. 2.8).

The name **peat** is used for organic accumulations composed of fibrous, semi-fibrous or amorphous materials. Unlike the other forms of organic matter, the plant remains are often still recognizable. Different forms of peat may be distinguished by their content of fibrous material. If a peat rich in fibrous material is rubbed, most of the plant fibres can be seen and it is referred to as **fibric**. Material in which the organic matter is almost completely

Fig. 2.7 Flocculated and dispersed soil. The flocculated soil is well-structured and has spaces (pores) through which air and water can move. These spaces are lost when the soil is dispersed.



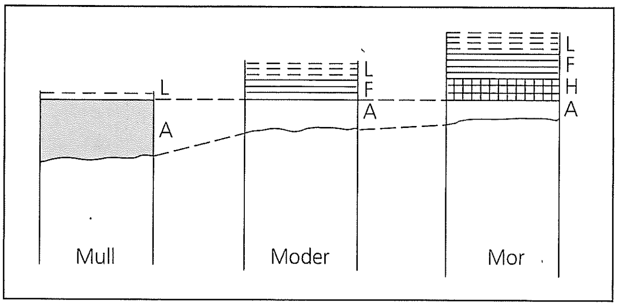


Fig. 2.8 Diagrammatic representation of the thin surface layers of organic matter (L, F and H). These make up the O horizon.

decomposed is called **terric**. (American pedologists use the term **sapric** for completely decomposed material, and that with an intermediate degree of decomposition is described as **hemic**.)

Mor develops beneath a heath or coniferous forest plant community and is associated with strongly acid soils. The plant litter is of low nutritive value and strong acidity severely restricts the range of soil fauna present. Breakdown of the litter is retarded so that three distinct organic horizons can be identified. These are the litter (L), fermentation (F) and humus (H) layers, which are included in the O horizon of the FAO horizon designation system. Much of the organic breakdown in the acid conditions of mor is achieved by fungi, and, as earthworms are usually absent from such soils, there is little incorporation of the humus, which rests abruptly on the mineral soil surface.

Organic matter can also accumulate in the B horizons of Podzols, forming a discrete layer separated from the surface mor and a very thin A horizon by the bleached albic E horizon that is characteristic of these soils. The mechanisms of podzolization will be discussed elsewhere, so a short explanation will suffice here. Organic compounds, often associated chemically with iron and aluminium, are moved down the profile to enrich the B horizon and give its upper part a dark colour, contrasting with the bleached eluvial horizon above and an iron-enriched spodic B horizon below.

Moder is a form of organic matter intermediate between mor and mull. It is often found under woodland conditions in association with Dystric Cambisols, and it is more acid and has a more restricted soil fauna than mull. It comprises decomposing organic matter and faecal pellets of the soil fauna, especially the springtails and mites. Litter (L) and fermentation (F) horizons are identifiable in approximately equal thicknesses. In the case of forms transitional to mull, an increased amount of humus incorporation into the mineral soil occurs as well.

Mull forms in freely drained, base-rich soils with good aeration. Such conditions are good for plant life as well and so there is a plentiful supply of plant litter, and associated with it a rich soil fauna including earthworms. The organic debris is completely broken down by the soil fauna and the humified remains are incorporated into the soil each year, so that none remains from one year to the next. Earthworms, in particular, are responsible for ingesting a mixture of plant and soil material and mixing it together intimately in their casts. In the mull form, the humus has a colloidal character and is intimately associated with the clay minerals, with which it forms the **clay–humus complex**.

The chemical composition of soil organic matter is extremely complex and is thought to be composed of high-molecular-mass substances which show no ordered structure when examined by X-ray diffraction. When analysed by conventional chemical methods, humus has fractions which are soluble in alkalis and in acids. A wide range of compounds and substances results from the activities of micro-organisms which break down the original plant material. These substances often have a phenolic structure with carboxyl groups (COOH) which dissociate to give a negative charge; the magnitude of this charge is dependent upon the pH and gives humus similar but larger cation exchange properties than clays, 200 $cmol_c\,kg^{-1}$. Some of these organic substances are capable of linking with metal ions as chelates, making them more soluble and easy to move in solution.

The clay–humus complex

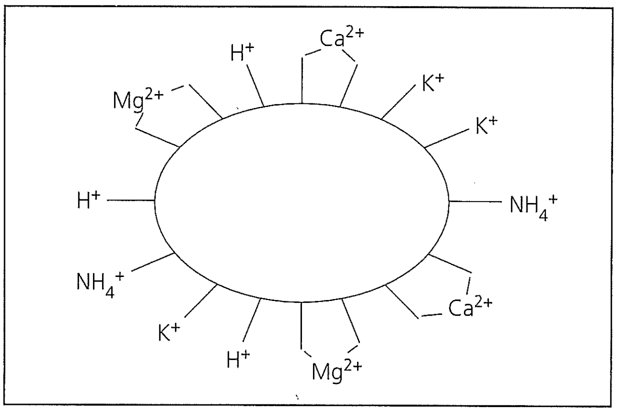
The chemistry of the soil is largely concerned with the chemical and physical activity of the minute particles of the clay–humus complex. As long ago as 1850, Thomas Way found that soils had the power to retain cations and that natural clays reacted in a similar manner to artificial silicates of lime and aluminium. Subsequent investigations have shown this piece of research to be correct, but the mechanism involved was not understood at the time. Acids, alkalis and their salts dissociate in solution to a certain degree – that is, they form ions with positive and negative charges. Pure water has the same property to a lesser degree, dissociating into hydrogen (H^+) and hydroxide ions (OH^-):

$$NaNO_3 = Na^+ + NO_3^-$$

$$H_2O = H^+ + OH^-$$

As a result of broken edges of the silicate clay crystals and ionic substitution within their structures, the clay mineral particles have a net negative charge. Similarly, the organic particles have a net negative charge derived from broken bonds of organic molecules. Thus the clay–humus particle effectively acts as a highly charged anion, which in colloidal chemistry is known as a **micelle** (Fig. 2.9). In the soil solution surrounding the micelle, attracted to it by the negative charge which they compensate, are numerous cations in a diffuse cloud, forming the Gouy layer. A molecular-thin layer of cations and anions at specific binding sites, known as the Stern layer, occurs at the clay–humus interface. Together, these two layers form the **electrical double layer** upon which models of the exchange system are built. The cations are said to be 'adsorbed' as they are capable of being displaced and exchanged for other cations.

Fig. 2.9 Diagrammatic representation of a clay–humus micelle with adsorbed ions of hydrogen (H^+), calcium (Ca^{2+}) and potassium (K^+) (after Eyre, 1969).



The total amount of exchangeable cations held by the clay–humus complex is the cation exchange capacity (CEC) of the soil, and the **base saturation** is the amount of base cations (Ca^{2+}, Mg^{2+}, K^+ and Na^+) present, expressed as a percentage of the total which includes hydrogen ions (H^+).

The exchange capacity of soils in the humid temperate regions tends to be dominated by the ions of hydrogen and calcium, together with lesser amounts of magnesium, sodium and potassium. In the humid tropics, strongly leached soils are dominated by hydrogen ions and the strong acidity allows aluminium ions (Al^{3+}) to come into solution, as also occurs in base-poor soils of humid temperate regions, exacerbated by the problems of acid rain.

With increasing dryness of climate, there is less leaching, and calcium and magnesium ions dominate the exchange positions of the clay–humus complex in central continental areas. In arid climates, where leaching is minimal and where the ground-water may be rich in soluble salts, sodium ions can come to dominate the exchange positions, resulting in soils of poor structure, in which virtually no commercial or subsistence crops can be grown. A clay–humus complex saturated by calcium and hydrogen ions has a flocculated state, forming stable soil crumbs. When sodium is the dominant cation, the clay and humus particles become dispersed and the soil is difficult to cultivate. Physiologically, plant growth is limited by the presence of salt in the soil because plants cannot draw moisture from the soil into their roots.

Soil structure

Soil structure is an important physical characteristic of any soil. In a soil, structure is produced by the individual particles of sand, silt and clay aggregating into larger units with specific shapes, known as **peds**. Most soils are **pedal**, that is they have structure. Those soils without structure, termed **apedal** soils, may be either **massive** or, in the case of coarse sandy soils, composed of unaggregated individual grains of sand, in which case the structure is referred to as **single grain**. Soil structure is encouraged by the presence of organic matter, and the gums and mucilages formed by bacterial breakdown of organic matter help to bind the peds together. The peds have been described as the 'architecture' of the soil, and the spaces around and through them act as channels to conduct water through the soil. These spaces between the peds, known as **pores**, are also important as they provide a habitat for the soil fauna.

It is not readily apparent from the surface that the total volume of air spaces in an organic-rich, medium-textured topsoil can be as high as 60 per cent, but is usually around 50 per cent. Cultivation reduces the number of larger pore spaces, which are important for the movement of air and water through the soil. An average figure of 45 per cent is quoted for nineteen cultivated Georgia (USA) soils, with 57 per cent pore space in neighbouring uncultivated soils.

Soil is readily classified into six structural types: structureless, platy, crumb, blocky, prismatic and columnar. Structure is described according to the type and size of structure and how well-formed the structure is in the soil. The different types are described in Table 2.2 and illustrated in Fig. 2.10.

Maintenance of soil structure is important for agriculturalists the world over, for unless a soil is well-structured, crop yields are depressed, and structureless soils are more liable to erosion. Soil structure can be weakened by over-cropping or over-grazing, both of which reduce the organic matter content to a level where collapse of structure occurs. The action of heavy farm machinery repeatedly passing over the soil can cause compaction and plough-pans in the subsoil. In the topsoil, structures may be crushed and the wheelings form channels which erosion rapidly exploits.

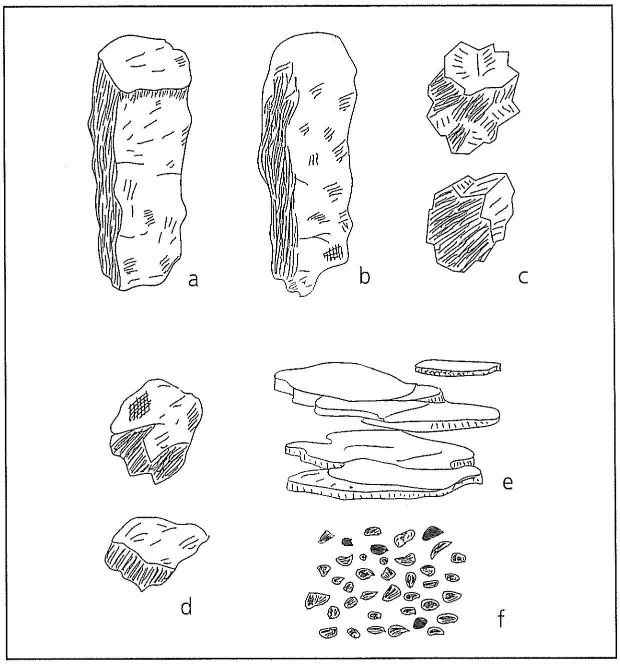


Fig. 2.10 Soil structures are formed by the aggregation of sand, silt and clay particles: (a) prismatic, (b) columnar, (c) angular blocky, (d) subangular blocky, (e) platy, (f) granular or crumb (after USDA, 1951).

Soil colour

One of the features a pedologist first looks at when describing the profile of a soil are the horizons present, mainly distinguished by changes in colour. The use of colour as a distinguishing feature for a soil horizon is a simple approach, but colour is indicative of many other, more complex soil properties. In many cases, soil colour is attributable either to the presence of organic matter or to the state of oxidation or hydration of the iron minerals present. Black and very dark brown colours are indicative of organic matter accumulation, and usually these colours are confined to the surface horizons. Grey colours are associated with removal or segregation of iron, especially in poorly drained conditions where they are accompanied by yellowish or reddish mottles. Reddish-brown colours are usually found in well-drained soils, and whitish colours are normally associated with accumulations of various salts in the soils of arid regions. Where there is strong summer drying of soil, as in areas of Mediterranean climate, soils develop bright red colours of iron in the hematite state. Colours of soils are described using the colour charts of the Munsell colour system which describe soil colours in terms of hue (the basic spectrum colour, e.g. red {R} or yellow {Y}), value (darkness or lightness from white to black) and chroma (the strength of the colour).

Soil air

The atmosphere penetrates into the soil along the pores and fissures. After rain, when excess moisture has drained from the soil, the volume of the remaining air-filled pores is referred to as the **air capacity**.

	Plate-like with one dimension (the vertical) limited and greatly less than other two; arranged around a horizontal plane; faces mostly horizontal	Prism-like with two dimensions (the horizontal) limited and considerably less than the vertical; arranged around a vertical line; vertical faces well defined; vertical angular.		Block-like; polyhedron-like, or spheroidal, with three dimensions of the same order of magnitude, arranged around a point.			
				Block-like; blocks or polyhedrons having plane or curved surfaces that are casts of the moulds formed by the surrounding peds		Spheroids or polyhedrons having plane or curved surfaces which have slight or no accommodation to the faces of surrounding peds	
		Without rounded caps	With rounded caps	Faces flattened; most vertices sharply angular	Mixed rounded and flattened faces with many rounded vertices	Relatively non-porous peds	Porous peds
Class	Platy	Prismatic	Columnar	(Angular) blocky[1]	Subangular blocky[2]	Granular	Crumb
Very fine or very thin	Very thin platy; <1 mm	Very fine prismatic; <10mm	Very fine columnar; <10mm	Very fine angular blocky; <5mm	Very fine subangular blocky; <5mm	Very fine granular; <1 mm	Very fine crumb; <1 mm
Fine or thin	Thin platy; 1 to 2mm	Fine prismatic; 10 to 20mm	Fine columnar; 10 to 20mm	Fine angular blocky; 5 to 10mm	Fine subangular blocky; 5 to 10mm	Fine granular; 1 to 2mm	Fine crumb; 1 to 2mm
Medium	Medium platy; 2 to 5mm	Medium prismatic; 20 to 50mm	Medium columnar; 20 to 50mm	Medium angular blocky; 10 to 20mm	Medium subangular blocky; 10 to 20mm	Medium granular; 2 to 5mm	Medium crumb; 2 to 5mm
Coarse or thick	Thick platy; 5 to 10mm	Coarse prismatic; 50 to 100mm	Coarse columnar; 50 to 100mm	Coarse angular blocky; 20 to 50mm	Coarse subangular blocky; 20 to 50mm	Coarse granular; 5 to 10mm	
Very coarse or very thick	Very thick platy; >10mm	Very coarse prismatic; >100mm	Very coarse columnar; >100mm	Very coarse angular blocky; >50mm	Very coarse subangular blocky; >50mm	Very coarse granular; >10mm	

[1] Sometimes called nut. The word 'angular' in the name can ordinarily be omitted.
[2] Sometimes called nuciform, nut, or subangular nut. Since the size connotation of these terms is a source of great confusion to many, they are not recommended.

Table 2.2 Types and classes of soil structures (after USDA, 1951)

Air within the soil is a natural continuation of the atmosphere above the soil, but although it is similar in some respects, it differs in others. Compared with atmospheric air, soil air is usually saturated with water vapour and it has a greater concentration of carbon dioxide.

The figures given in Table 2.3 indicate that the soil air has slightly less oxygen and more carbon dioxide (eight times) than atmospheric air, but the amounts of both gases vary considerably according to the activity of the micro-organisms living in the soil. Addition of leaf litter or organic manure greatly stimulates the soil fauna, which may result in the depletion of oxygen and an increase in carbon dioxide. The exchange of these gases with the atmosphere above the soil is hindered if the soil pores and fissures are small or limited in number. When the pores are filled with water, fresh oxygen cannot easily diffuse in, and such oxygen that may be present in the soil is soon consumed so that anaerobic conditions develop. It is in these conditions that the growth of most plants is inhibited and conditions favouring chemical reduction occur in soils.

Considerable energy is currently being exerted by the scientific community in ascertaining the amounts and rates of change of the so-called 'greenhouse gases' in the atmosphere. As the soil holds large reserves of carbon (about twice the amount held in the vegetation), investigations are taking place to find the potential amounts of carbon dioxide and methane which might be released from soils with changing land-use and climatic conditions. From these investigations it has emerged that soils are an important factor influencing the amount of these trace gases. Specifically, the actions of deforestation and cultivation interrupt the natural cycling of carbon, causing the release of carbon dioxide. Cultivation of rice results in the release of methane from the flooded, anaerobic soils of paddy fields, mainly through the stems of the rice plants. Denitrifying bacteria release nitrogen into the atmosphere, and nitrogenous fertilizers added to arable soils increase emissions of nitrous oxide from soils to the atmosphere.

Table 2.3 Average composition of soil air (per cent by volume)

	Oxygen	Carbon dioxide	Nitrogen
Soil air	20.65	0.25	79.20
Atmospheric air	20.97	0.03	79.00

The movement of gases into and out of the soil takes place by diffusion, a process which depends upon the differences in concentration of gases between the soil atmosphere and the free atmosphere above the soil. Even in freely drained soils with wide pores, some water is trapped by constrictions, so the process is not straightforward. Changes in atmospheric pressure can cause the soil to 'breathe' and the gustiness of the wind may assist gaseous movement. The passage of water through the soil may also displace air from below the wetting front as it moves downwards.

Oxygen diffuses into the soil, where it is taken up by roots and used in microbial activity. During the summer, the demand for oxygen is between 7 and $35\,g\,m^2\,d^{-1}$ and unless there is a saturated layer, the process of diffusion is sufficiently rapid to meet the demand. Conversely, as a result of biological respiration, carbon dioxide diffuses from the soil pores out into the atmosphere at a rate of between 1.5 and $6.7\,g\,m^2\,d^{-1}$ (but some also dissolves in water), and it has been predicted that, other than in anaerobic pockets, the concentration of carbon dioxide in soil air should not normally exceed 1 per cent.

Soil water

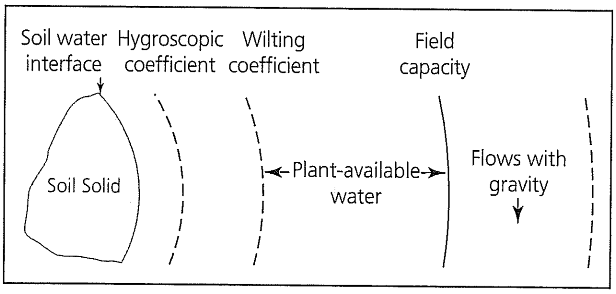
The presence of air and water in the soil is almost complementary, for if the soil is saturated with water the air is driven out. In a poorly drained, saturated soil, almost all the pore space between the peds is occupied by water. If the soil is allowed to drain so that the water contained in the larger pores and cavities is removed, the water which is lost is known as **gravitational water**. About two days after heavy rain or irrigation, a freely drained soil has lost all its gravitational water and is said to be at **field capacity**. In this state, considerable amounts of water remain in the finer pores, held by capillary attraction in pores and attached to the surface of soil mineral particles.

If the soil continues to lose moisture from the reserves of capillary water, the point is eventually reached at which plants wilt because they cannot obtain sufficient moisture to maintain turgor. This **wilting point** is not fixed, for plants vary greatly in their capacity to extract water from the soil. The amount of water held between the field capacity and the wilting point of a soil is referred to as the **available water capacity** (Fig. 2.11). This amount will vary according to the texture, organic matter content and structure of the soil, and it is vitally important when considering the ability of a soil to supply a growing crop with its moisture requirements.

Further amounts of moisture can be extracted from the soil in the laboratory. It is possible to bring the soil to air-dry conditions, but as this is rather a variable state depending upon the humidity of the atmosphere, an oven-dry basis, 105°C, is used for most laboratory determinations. Water held at temperatures above oven-dry is referred to as hygroscopic water and is largely unavailable for plants. This water is virtually part of the mineral structure of the soil and may occur as a coating on mineral grains with a thickness of only a few molecules.

The moisture regime of many soils that have ground-water can be simply demonstrated by drilling 10 cm diameter auger holes to different depths, lining the holes with porous tiles and allowing the water level to come to equilibrium. Observations made at suitable intervals throughout the year will reveal the variation of the water level. Infiltration (the rapidity with which water can enter into the soil) can be measured using a ring infiltrometer in which the time taken for a known volume of water to infiltrate is recorded. Soil permeability can also be measured in an auger hole by the rapidity with which the water level returns to normal following removal of water. In the laboratory, soil permeability may be measured by an instrument called a permeameter, in which the passage of water through a soil sample is measured in either unsaturated or saturated conditions.

Fig. 2.11 The width of a film of moisture around a soil particle determines the tension at which it is held. At the outer limit water flows with gravity, but as drying narrows the water film it is more and more tightly held. Water is available for plants between field capacity and the wilting coefficient.



In order to bring some uniformity to the measurement of soil moisture contents, a system was developed that measures the suction required to pull the water out of the soil. Originally expressed in terms of the tension exerted by the length of a column of water, the range of the numbers necessitated using a logarithmic scale (pF). At saturation, the length of the column is nil and so the suction pressure (in bars) exerted is 0.0 and the pF is 0.0. At field capacity, the suction pressure (–0.33 bar) is represented by a water column 333 cm long or pF 2.5, and at wilting point the suction pressure (–15 bars) is theoretically a water column of 15,000 cm or pF 4.2. However, in recent years the forces which facilitate the movement of water in soils have been collectively referred to by their potential to cause movement, and their measurement is in kilopascals (kPa). The potential of soil water (Φ) is a measure of the amount by which the free energy of water in the soil is reduced by gravitational, osmotic (plant) and matrix (capillary) potentials, which together make the total potential force available for water movement.

Simply stated, water may move through soils in three ways: it may flow freely through the wider fissures and pores, but it also responds to capillary forces and the osmotic pull exerted by plants. Water may also move in the vapour phase under the influence of a temperature gradient, a phenomenon which can be observed in soils of arid areas.

The soil solution

Any soluble constituents present in the soil will dissolve in the water contained in the soil pores and contribute to the **soil solution**. This is the medium through which plants are supplied with the nutrients they require. As has already been observed, inorganic salts dissociate in solution into ions. Many of these ions are attracted to and adsorbed by the clay–humus micelles, but an equilibrium is reached between the ions occupying the exchange positions and those in the soil solution.

In the case of hydrogen ions, their concentration in solution is indicated by the pH scale, in which neutrality is at pH 7. Values below pH 7 are acid and those above pH 7 are alkaline. The general pH range of soils is from pH 3 to pH 10. In humid regions the normal range is from pH 5 to pH 7 and in arid regions from pH 7 to pH 9. Extremely acid conditions, pH 2, may be found in some acid sulphate soils (Thionic Fluvisols), whereas at the other end of the scale, some strongly alkaline soils (Solonetz) may reach a value of pH 10. Soil pH represents an easily determined soil characteristic, which also has a general usefulness. Plant nutrients become less available at the extremes of pH and other elements become available in toxic amounts, so the pH value is often a good guide in the diagnosis of fertility problems.

A significant role of the soil solution is to transport soil constituents from one horizon to another. Movement may be either in solution or suspension; soil constituents will be taken into true solution if they are soluble, or small particles of silt, clay or organic matter may be washed down the profile in suspension. Movement of chemical constituents is also facilitated by chemical linkage (chelation) such as with iron and organic matter, when the iron is mobilized by its association with organic molecules.

In humid climates on level ground, the movement of the soil solution is vertically downwards from the upper eluvial horizons to the lower illuvial horizons of the profile. On sloping sites, soil moisture also moves downslope (laterally through the soil profile), and over a long period of time can produce the sequence of different soils on slopes known as a **catena**. In arid climates, especially where the ground-water table is at shallow depth, movement of soil moisture can be upwards as evaporation draws moisture to the soil surface. Any soluble salts will then be precipitated in the upper soil horizons, where they cause problems for most agricultural crops.

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3 Factors of soil formation

The question of how a soil forms inevitably leads first to a consideration of the environmental factors which control the processes of soil formation. Just over 100 years ago, the famous Russian soil scientist Dokuchaev suggested five 'soil formers' which act together to set the limits of soil formation at any one place. These were: parent material, climate, age of the land, plant and animal organisms, and topography. Simply expressed, the concept stated that in the process of soil formation, the parent material is acted upon by the climate and the organisms, over a period of time signified by the age of the land. Topography was also considered, as it has much to do with soil–water relationships and the manner in which gravity can affect soil formation.

Fifty-five years ago, an American soil scientist, Hans Jenny, considered a number of similar 'soil-forming factors', and went on to show how they were functionally related in the form of an equation:

$$s = f'(cl, o, r, p, t, \ldots)$$

in which *s*, a soil property, is dependent upon (or is a function of) the soil-forming factors of climate (*cl*), organisms (*o*), relief (*r*), parent material (*p*), and time (*t*). The three dots imply any other significant soil-forming factors, such as the influence of human beings on soil formation.

The usefulness of this approach is that it is possible to take each soil-forming factor in turn and consider its variation or effects against the others. Whilst the approach is only semi-quantitative, it has the advantage of indicating the relationships of the soil-forming factors, bringing them together for the purposes of discussion. In view of the importance of the factors of soil formation in a consideration of the geography of soils, a brief discussion of their individual roles will be considered.

Climate

Climate is a composite concept which includes the temperature, rainfall, humidity, evapotranspiration, duration of sunshine and many other atmospheric variables. The temperature experienced at any one place sets the speed of any chemical reaction in the soil, and the rainfall percolating through the soil provides a medium in which chemical reactions can take place, as well as providing a mechanism for the movement of soil constituents.

There is a considerable range for both temperature and rainfall throughout the world: rainfall, for example, varies from less than 100 mm in deserts to more than 12,500 mm in the wettest areas, and temperatures that have an annual range of only 0.5 or 1.0°C contrast with those having a range of 43°C. The annual rainfall figures themselves are not always a good indication of the type of soil which will be formed. The total amount of precipitation is significant, but it is as important to assess the effect of seasonal rainfall, and whether it accompanies the cool or the warm season. The intensity of the rainfall is also important, as the effect of short, intense downpours can be very different to long periods of gentle drizzle. These features, as well as the local micro-climate, are important considerations, but unfortunately they are not often recorded.

The rainfall which eventually penetrates into the soil is less than the recorded rainfall, as moisture is diverted through runoff and evaporation from the soil and vegetation surfaces. Plant roots take up water from the soil and, after passing through the plants, it is returned to the atmosphere by the process of evapotranspiration (Fig. 3.1). Water that is not intercepted by plant roots percolates to greater depths and either emerges again in seepages further downslope or joins the ground-water.

Temperature may not seem to have such an

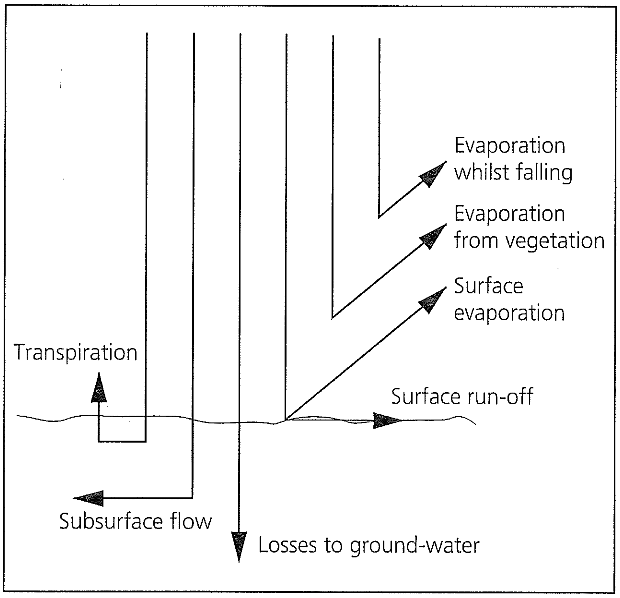


Fig. 3.1 Water entering the soil is not the same as the rainfall; the diagram shows losses incurred before the water reaches the soil. Some water may also be received by downslope movement through the soil.

important role to play in soil formation, but following an idea of Ramann (1928), it is possible to show how it can be significant over a long period of time. In Table 3.1 the relative dissociation of water (into hydrogen and hydroxyl ions) is taken as an index of the rate of chemical activity, and this, multiplied by the length of the weathering period, gives a **weathering factor**. It can be seen that in tropical regions the effectiveness of weathering is almost ten times that in arctic regions and three times that in temperate regions. Weathering has also been in operation for much longer in tropical regions, where it has not been interrupted by a change in climate as during the glacial periods of higher latitudes. Deeper weathering is therefore a characteristic feature of many soils in tropical regions. Up to 50 m of weathered mantle (regolith) may occur, though the depth is very variable. In temperate regions, the depth to unaltered geological materials may be only one metre or less.

A number of attempts have been made to reduce the effects of climate to a single figure or climatic index. Each has its merits, but most attempts lack a worldwide significance. For example, Thornthwaite's precipitation:effectiveness index works well in North America, but is less accurate in its assessment of the climate of Europe. It is possible to show correlations with rainfall for a number of different soil characteristics. One obvious example is the leaching of calcium carbonate from the soil: the greater the rainfall, the deeper the horizon of calcium carbonate appears in the soil, until it is completely leached out (Fig. 3.2). The formation of clay mineral type and content is also weakly related to the amount of rainfall. Correlations show an increase of clay content accompanying an increase in temperature, and the effect of a higher temperature is also seen in the more rapid decomposition of organic matter in soils of tropical areas.

Unlike the older, climatically based systems of classification, the pedological classification of soils used in this book does not completely mirror the pattern of climate on the Earth's surface. Accordingly, eight climatic zones have been distinguished on the basis of length of growing period, the length of frost-free season and the incidence of rainfall. These 'agro-ecological zones' are now commonly used by scientists involved with soils and agriculture at an international level in the Food and Agriculture Organization of the United Nations (FAO), the Consultative Group of International Agricultural Research (CGIAR), the International Board for Soil Research Management (IBSRAM) and the International Soil Reference and Information Centre (ISRIC).

The length of growing period is estimated from

Table 3.1 Ramann's weathering factor (after Jenny, 1941)

	Average soil temperature	Relative dissociation of water	Days weathering	Weathering factor	
				Absolute	Relative
Arctic	10	1.7	100	170	1
Temperate	18	2.4	200	480	2.8
Tropical	34	4.5	360	1620	9.5

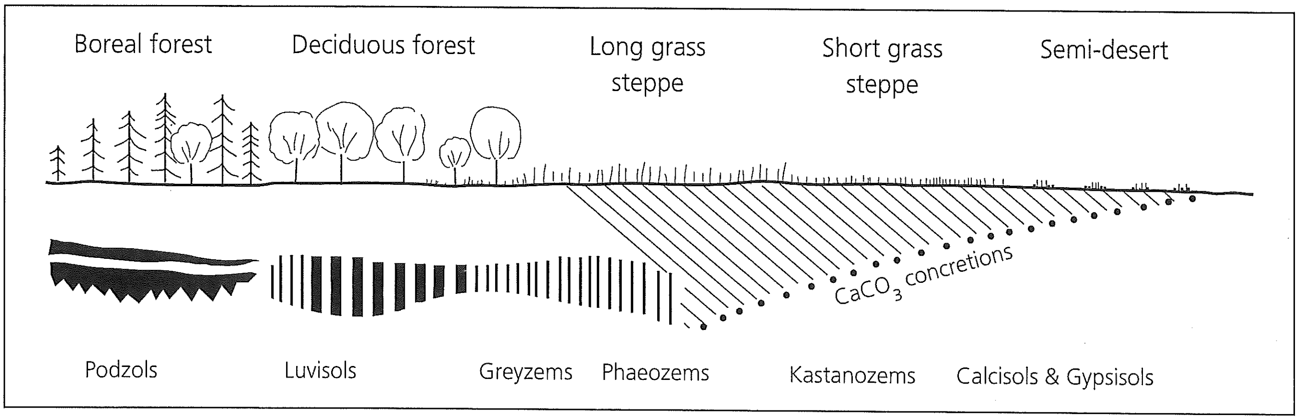


Fig. 3.2 The amount of rainfall can broadly be correlated with the depth to which $CaCO_3$ is leached. Thus it is completely removed from Podzol but can appear at the surface of a Calcisol.

the number of days in the year when precipitation exceeds half the evapotranspiration plus a period of time sufficient to remove 100 mm of water from excess water stored in the soil and minus the number of days when the daily mean temperature is less than 6.5°C. The period of time remaining reflects the time when both water and temperature are favourable for crop growth. A period of less than 75 days is used to define an arid climate, between 75 and 270 a seasonally dry climate, and more than 270 days a humid climate.

The frost-free period is defined as the number of days during the year when the average of the extreme minimum temperatures is above 2°C. Using these definitions, the boreal and polar zones have a frost-free period of less than 75 days, the cold zone has a frost-free period of between 75 and 135 days, and the temperate climate has more than 135 days with the average of the six warmest months below 25°C. Within the tropical regions, the seasonally dry sub-tropics are distinguished from the humid tropics by a dry season which extends from 90 to 285 days, whereas the humid tropics have favourable growing conditions all year round.

Organisms

The role of plant and animal organisms in soil formation is of critical importance, for without life there can be no true soil. It has become customary to subdivide the soil organisms into two groups based on size, micro-organisms and macro-fauna, each of which has a particular part to play in soil formation. The green plants synthesize carbohydrates, starches, proteins and other compounds from simple inorganic substances derived from the soil and air by the process of photosynthesis, using energy provided by the sun. When the plants die and the leaves fall to the ground, a rich source of food and energy is provided for the organisms which live in the soil and which are responsible for converting plant debris into the amorphous substance called humus.

Micro-organisms

The numbers of soil-living micro-organisms can be very great indeed (Table 3.2). However, numbers are not necessarily a good guide to their activity, as many organisms may be in a dormant state when sampling occurs; their activity is best measured indirectly by the amount of carbon dioxide evolved. Soil micro-organisms are members of both plant and animal kingdoms and include viruses, bacteria, actinomycetes, protozoa, fungi and algae.

The biomass of bacteria in the soil can amount to a live weight of 3360 kg per hectare (3000 lb per acre). Bacteria live in the thin water films surrounding soil particles and usually have organs of locomotion (cilia or flagella), which enable them to move through the water. Both aerobic and anaerobic bacteria are present in soils, but some normally aerobic bacteria (faculative anaerobes) can also live without oxygen. Aerobic bacteria can rapidly deplete a saturated soil of oxygen, causing chemical reducing conditions. Most soil bacteria

	Broadbalk Field		
	Manured	Unmanured	Grassland
Insects			
Springtails	40.6	28.3	54.0
Beetle larvae	5.9	0.9	2.3
Fly larvae	19.4	3.8	11.1
All others	3.4	0.5	11.0
Myriapods	4.5	1.8	1.8
Arachnids			
Mites	6.5	1.9	2.9
Spiders	0.17	0.07	1.2
Woodlice	0.04	0.05	–
Gastropods	0.04	–	0.05
Oligochaetes	2.6	0.6	8.4
Nematodes	1.5	0.2	7.6
Totals	84.6	33.2	100.5

Table 3.2 Numbers of small animals found in soil at Rothamsted (1936–1937). Numbers represent millions per acre in top 22 cm (9 inches) (after Russell, 1950)

obtain their energy from organic matter, and in the process produce the substance called humus. Chemotrophic bacteria can make use of inorganic substances; some of these are capable of breaking down sulphur or cyanide compounds as well as complex herbicides and pesticides to obtain their energy requirements. Soil bacteria are preyed upon by protozoans, the most numerous of which belong to the amoebae and the flagellate classes. Both free-living (*Azotobacter*) and symbiotic nodular bacteria (*Rhizobium*) are vitally important for extracting nitrogen from the air, providing a natural source of this essential plant nutrient. Certain algae of the class Chlorophyceae can also fix nitrogen. Less desirable are the denitrifying bacteria (*Pseudomonas*) which reduce nitrate to nitrogen or nitrous oxide, particularly in poorly drained soils.

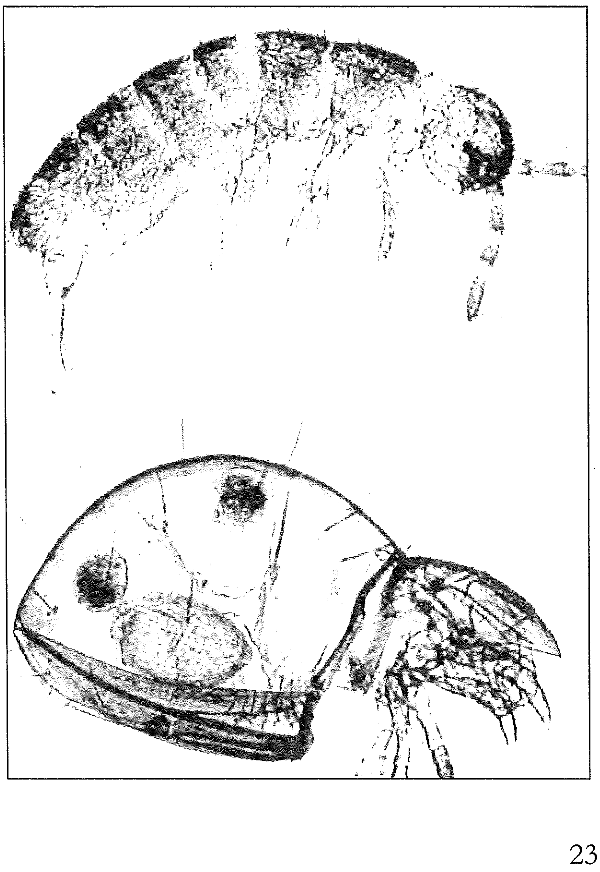
Of equal importance to the bacteria are the soil-living fungi which can amount to up to 0.2 per cent of the soil volume. Mushrooms and toadstools are the obvious above-soil evidence of the mass of fungal hyphae which ramifies through any decomposable substance within the soil. In some circumstances, such as very strong acidity, fungi are more important than bacteria in the breakdown of organic debris. Organisms with superficial similarities to fungi which occur in neutral or alkaline conditions are the actinomycetes, from which the Nobel-prize-winning soil microbiologist, Waksman, first extracted the antibiotic streptomycin.

Macro-fauna

Many different animals live in the soil; small ones live in the spaces between the soil structures, the larger ones can create their own burrows. In this category, animals belonging to the Arthropoda are the greatest in number, mostly living within the top few centimetres of the soil. In the temperate regions, the springtails (Collembolae) and the mites (Arachnidae) are the most populous (Fig. 3.3). The springtails are responsible for breaking organic matter down into small fragments; many of the mites are carnivorous, feeding upon the springtails.

The earthworms (Oligochaeta), the subject of studies by Darwin and the famous English naturalist Gilbert White, play an important role in the mixing of organic matter with mineral matter. *Lumbricus terrestris*, for example, can be seen to

Fig. 3.3 Soil fauna: examples of a collembola (top) and a mite (from Russell, 1961).



actively draw fallen leaves into its burrow. As it feeds, the earthworm ingests a mixture of soil and organic matter which is intimately mixed in the worm's digestive system; the resulting wormcast is a close mixture of humus and soil, slightly alkaline and richer in plant nutrients than the surrounding soil which has not passed through the earthworm. Casting on the soil surface by earthworms eventually leads to the development of a deep, organic-rich topsoil at the A horizon which is gradually thickened. The activities of earthworms – burying organic matter, burrowing, casting and soil mixing – are responsible for the production of the humus form known as 'mull'. Only *Allolobophora longa* and *Allolobophora nocturna* cast on the surface; the other species cast in cavities below the soil surface. Cultivation, which kills about 25 per cent of earthworms, and the lower amount of plant litter associated with arable land have the effect of reducing the earthworm population.

In addition to his evolutionary studies, Charles Darwin (1809–1882) was interested in the activities of earthworms. He measured the weight of worm-casts and calculated that they amounted to 10 tons per acre per year, and that in 50 years the whole volume of soil down to 9 inches (23 cm) would have been brought to the surface by the worms. To measure the effect, Darwin spread chalk and ashes on the surface of the soil in 1842 and by 1871 observed that these were buried by a layer of 7 inches (17.5 cm) of soil. When a later investigator checked the same experimental plot in 1942, the position of the layers of chalk and ashes had not changed.

Whilst the role of the earthworms is beneficial, the same cannot be claimed for the nematodes. This group of creatures is parasitic on plants and animals, and those that live in the soil are no exception. Many nematodes attack plants and reduce yields, as well as transmitting viruses of plant diseases. During drought or in the absence of a host, nematodes are capable of forming a cyst in which the eggs are protected until conditions improve. The species *Heterodera rostochiensis*, the potato root eelworm, is stimulated by exudates from growing potato roots to break out of its cyst and to infect the new plants of the following crop.

Many arthropods spend all or some part of their life cycle in the soil. Centipedes (Chilopoda) are active predators on other soil fauna, but the food of millipedes (Diplopoda) is decaying plant litter, a diet which they share with the woodlice (Isopoda). Both reduce the plant litter to very small particles which are then attacked more readily by micro-organisms. Crop pests such as leather jackets (Diptera) and wireworms (Coleoptera) are larval forms which live in the soil and feed on plant roots. Some insects lay their eggs in the soil and, when hatched, the larvae enter the plants, returning to the soil to pupate before emerging as the adult form.

The colonial insects, ants (Hymenoptera) and termites (Isoptera), are both important inhabitants of the soil. The termites are restricted to tropical areas but ants occur widely in the soils of temperate areas. Both species create their own sheltered environments within the soil, the termites constructing mounds several metres high, with galleries penetrating deep into the subsoil. Ants are predacious, but termites bring plant material into their nests which they 'cultivate' in their underground fungus gardens. It has been estimated that termites are capable of transferring over 1 tonne per hectare per annum ($t\,ha^{-1}\,a^{-1}$) of subsoil to the surface of the soil.

Slugs and snails (Gastropoda) feed mainly on organic debris, which they break down and make more readily available for bacterial colonization. In moist, clayey soils, slugs living just below the soil surface can be a pest, eating new shoots of seeds before they emerge from the soil.

Small mammals, including moles, shrews, rabbits and gophers, and some larger animals, such as foxes and badgers, make their homes in the soil. Their role is restricted to loosening and mixing the soil through their activities. Some birds, too, may influence the nature of the soil. Certain species nest below ground, and small stones from crow's crops have been observed to accumulate below popular roosts in Wisconsin USA.

Larger mammals may be responsible for some soil disturbance such as at salt-licks or mud-wallows, and grazing herbivores provide quantities of dung which contribute to the soil organic matter content. The activities of human beings, however, have the most profound effects on soils and these are considered in Chapter 8.

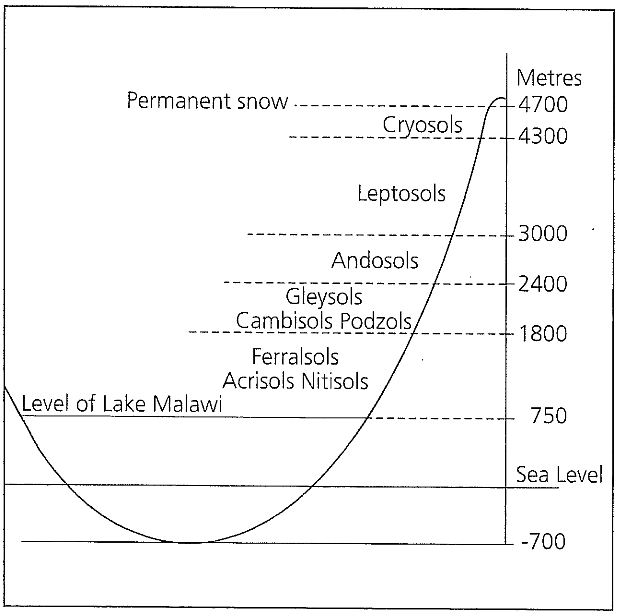
Plants

Although vegetation is ultimately dependent upon the climate, it can also function as an independent variable. A good example occurs where an oakwood is replaced by a coniferous plantation: the demands made on the soil are changed and the type of litter supplied to the soil is different, leading to increased podzolization. Where oakwood has colonized Chernozem soils, leaching is encouraged and the soils become degraded.

The regular fall of litter to the soil surface from the vegetation, together with decaying roots, supplies the soil with its organic matter and many of the soil fauna with their nutrient supply. Inputs of litter from forests range from about 2 t ha^{-1} a^{-1} in temperate regions to 10 t ha^{-1} a^{-1} in the tropical rain forest. In an old grassland the annual input of carbon (as roots) was between 2.0 and 2.5 t ha^{-1}. The supply of organic matter to cultivated land from roots is in the order of 0.3 to 1 t ha^{-1} of carbon for small grains, but amounts are dependent upon the quantity of fertilizer applied.

In seasonal environments, the soil microbial population responds rapidly to the availability of food, increasing in number until frost or other adverse conditions inhibit their activity. The activity of the soil flora and fauna converts the plant debris into humus, producing the forms of soil organic matter known as mull, moder or mor, described in Chapter 2.

Fig. 3.4 Altitudinal sequence of soils from Lake Malawi to the snowline.



A most significant zone of interaction between plants and the soil is in the surroundings of roots, sometimes referred to as the **rhizosphere**. The presence of roots provides conditions that encourage the growth of the microbial population. Debris sloughed off plant roots provides food for free-living forms of micro-organisms, and exudates from the roots add to the nutrient supply. Mycorrhizae, symbiotic associations between fungi and plant roots, enable many plants to gain easier access to nutrients. The rhizosphere organisms include nitrogen fixers; their contribution to the nitrogen budget of the soil is described as 'uncertain but significant' and may amount to up to 39 kg ha^{-1} a^{-1}.

Relief

The well-known effects of relief on climate include depression of temperature and increase in rainfall. As an important soil-forming factor, increasing elevation combines with climate to give a succession of different soil-forming conditions, leading to changes in the soil type produced. The succession of soil zones on the slopes of high tropical mountains exhibits this soil pattern very well. A sequence from Ferrallitic Soils on the shore of Lake Malawi to soils at the snowline is given in Fig. 3.4.

Also associated with upland areas is a greater incidence of cloudiness and hence less solar warming; evapotranspiration is also reduced, leading to colder, wetter soil conditions which slow plant decomposition and favour the development of thicker organic horizons. Such conditions are particularly prevalent, for example, in the northern and western uplands of the British Isles.

The compass direction in which a slope faces is described as its **aspect**; the aspect of a slope can greatly affect the amount of solar warming received, especially in the temperate climate zones. Differences in soils will develop in response to the different conditions on sun-facing compared with shaded slopes (Fig. 3.5).

Slope also influences the movement of water on the land and through soils. Sites on the crests of hills and steep slopes actively shed water into low-lying areas which receive proportionately more water for

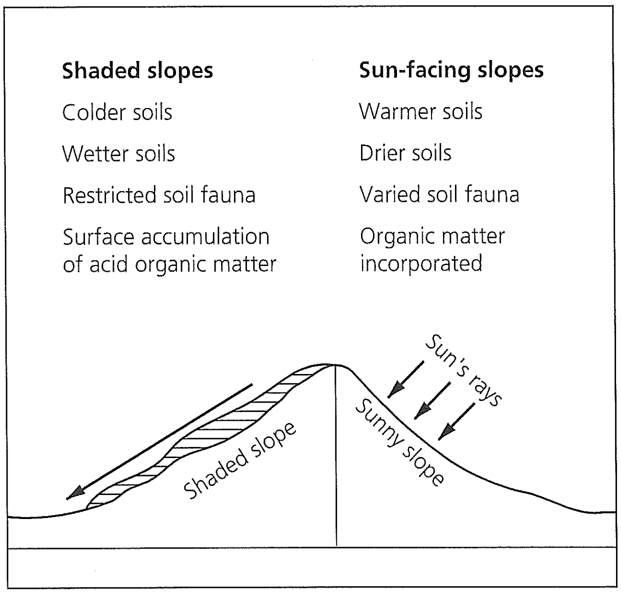
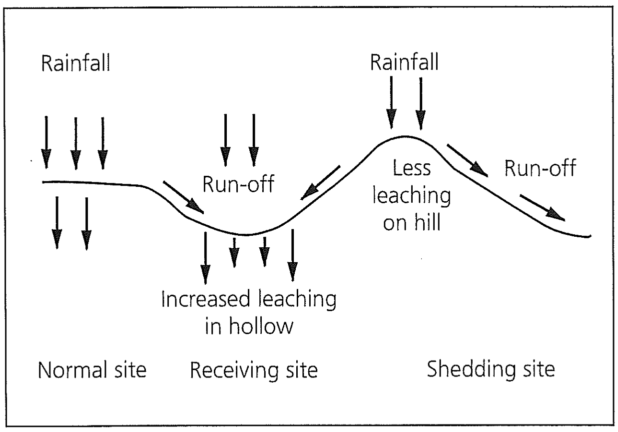


Fig. 3.5 The effect of aspect and relief on soils situated on shady and sunny slopes.

percolation through the soil profile. Consequently, lower slopes remain wetter for longer periods as water drains from soils higher upslope. Level sites are described as 'normal', although most sites have some slope, even if it is extremely slight (Fig. 3.6).

Gravity acts upon slopes with a vector which exerts its influence in the direction of the slope. There is a tendency for soil material always to be drawn in a downslope direction, whether it is in solution or in the solid state. The action of water, with the mobilization, transport and redeposition of soil materials by overland flow, throughflow and seepage, always works in a downslope direction.

Fig. 3.6 The effect of slope on water availability for soil formation.



Each wetting–drying cycle and freeze–thaw cycle also imperceptibly moves material downslope in the process known as creep. The long-term effects are shown in the nature of slopes and the inter-linked soils formed upon them. This relationship of a linked sequence of soils with position on the slope is known as a **catena**.

The word 'catena' relates to the way links of a chain, suspended at each end but left to hang freely, form a loop. The soils on a hillside are like the links forming the loop, arranged in sequence from interfluve crest to valley floor. The soil relationships of the catena were first identified by Milne in the 1930s in East Africa, and the concept has since proved very useful in the interpretation of soil–landscape relationships in virtually all environments. The catena was the basis for the development of a nine-unit landsurface model, which is helpful in understanding the linkage between processes which operate on slopes and the nature of the soils which develop. The nine landsurface units and their diagnostic properties are summarized in Fig. 3.7. The processes include the effects of gravity and water on the mobilization, transport and redeposition of soil materials by overland flow, throughflow, streamflow and mass movements. This pedogeomorphological approach brings together many aspects of Earth surface processes relevant to pedologists, and is sometimes referred to as a soil–water–gravity model.

The shape of the landscape also influences the movement of weathered debris. Around a concave valley head, creep will tend to concentrate potential soil parent material and soils will be deeper, whereas around the convex nose of a spur of higher land it will tend to be dispersed outwards over a wider area of lower land. As a general rule, soils will be deeper on footslopes where downslope creep has resulted in the accumulation of weathered material and eroded soil. These footslope accumulations are referred to as **colluvium**.

Certain American authors have found that successful prediction of soil attributes can be made using digital terrain models, enabling a link to be made between topography and quantifiable soil characteristics. A strong correlation has been observed between measured soil attributes, slope gradient and a 'wetness index'.

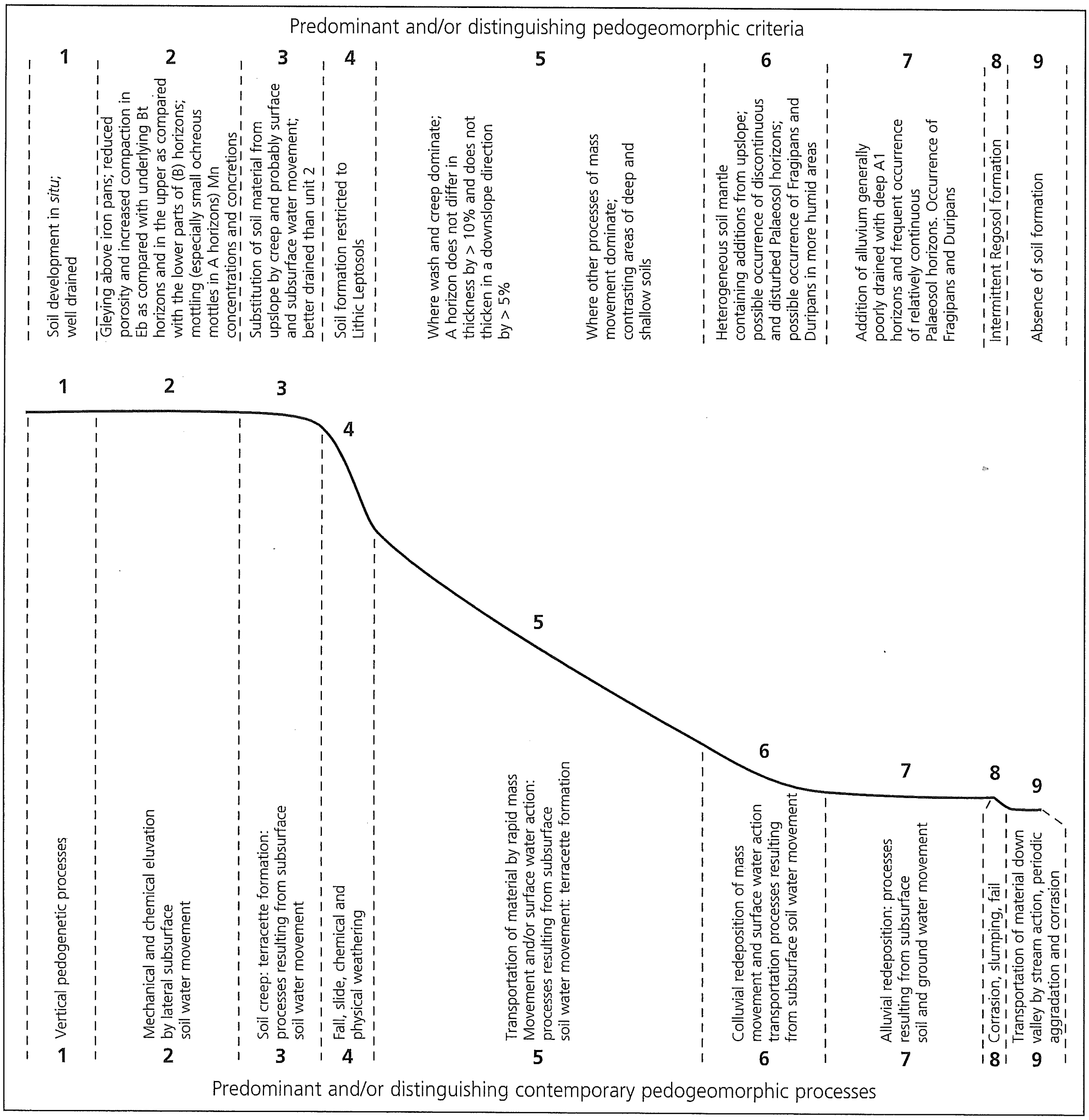


Fig. 3.7 The nine-unit landsurface model (after Conacher and Dalrymple, 1977).

Parent material

The parent material was described by Jenny (1941) as 'the initial state of the soil system', but it is more generally described as the consolidated or unconsolidated material, little affected by the present weathering cycle, from which the soil has developed. The more simple definition, that it is the material which lies beneath the true soil horizons, could be misleading, as many soils have developed from composite parent materials of differing origins and not simply from the rock which happens to lie at some depth beneath. For example, a soil can develop in a thick glacial till or loess layer overlying an almost unweathered rock; it is possible that there may be thin layers of both glacial drift and loess present in which the soil has developed, or the soil may be developed in both the superficial layers and the bedrock beneath.

It is useful at this point to distinguish between weathering and soil formation; the former is a geological process producing the **regolith**, and the latter is a pedological process, often working on and changing further the results of weathering into soils. Briefly, weathering involves the physical disintegration of the rocks, and the geochemical processes of solution, hydrolysis, carbonization, oxidation and reduction, as well as the rearrangement of the structure of some clay minerals and the formation of others. The soil-forming processes are considered in more detail in Chapter 4.

Weathering and soil formation can proceed either separately or together. In old, deep soils, such as occur in many inter-tropical areas, weathering will occur at depth, well below the genetic soil horizons which have formed near the surface. In younger soils, which are usually much shallower, the two processes may operate almost simultaneously within the same few centimetres of soil.

In most environments, the balance of soil-forming factors is such that different rocks, when weathered, will produce a regolith which in turn will give rise to different soil types, depending on the inherent mineral composition. This has been demonstrated in Scotland, where Cambisols form on basic igneous rocks and Podzols on acid igneous rocks under identical climatic conditions. The differences between calcareous and non-calcareous parent materials are of sufficient importance to be reflected in many soil classifications. The presence of large reserves of bases, in particular calcium and magnesium, maintains the iron, aluminium and humic constituents in a flocculated state, thus inhibiting movement and so retarding the development of horizons within the soil profile. In Wales, outcrops of the Carboniferous Limestone on Black Mountain, Dyfed, give rise to Cambisols at an elevation and under a rainfall which ensure Humic Gleysols on adjacent non-calcareous parent materials. Similarly, in the Peak District of Derbyshire, England, the limestones have a cover of Cambisols whereas the surrounding sandstones and shales only support moorland with Humic Gleysols and Gleyic Podzols.

Extremely porous, quartzose, sandy parent materials rapidly achieve a mature soil profile because they contain few soluble materials, and any constituents in solution or suspension can move easily through the profile and out of the soil. Thus many coarse, sandy parent materials throughout the geological succession develop Podzols upon their weathered mantles. Some of the most striking Podzols are found on the coarse-textured glacio-fluvial sands and gravels of former glaciated areas as well as on the sandstones of the Tertiary formations. Cambic Podzols, and Dystric and Chromic Cambisols are commonly found on steeply sloping sites, for example in Wales and Scotland, where they are developed from weathered debris of Lower Carboniferous strata.

In contrast, clay soils are usually slow-draining and slower to develop a fully mature profile than freely drained soils. Wide expanses of till plains in the lowlands of England have fine-textured Eutric Gleysols or Gleyic Luvisols (Surface-water Gley Soils); similar parent materials in the uplands develop peatyness and typical soils are Dystric Gleysols (Peaty Gleys).

Frequently, alluvial materials are sorted into layers with different particle sizes by fluvial action. They are often silty and possess organic matter in variable amounts and to greater depths than other soils. A combination of alluvial materials together with a covering of neutral or calcareous peat can provide highly fertile Terric Histosols (Earthy Peat Soils) when drained. Unfortunately, the action of draining and cultivation causes oxidation of the peat and most of the former peats of the Fens of eastern England have now disappeared, leaving humose silty mineral soils.

In the sub-humid and semi-arid lands, conditions are suitable for the development of soils rich in the clay mineral montmorillonite (smectite). Leaching of soil constituents from the upland areas to the lowlands provides the raw materials for recombination of elements into new minerals. Along the valley floors of the savanna lands, black, montmorillonite-rich clays have formed. These dark-coloured, deeply cracking clayey soils are discussed under the heading of Vertisols in Chapter 6.

Where blown sand, loess or volcanic ash accumulates, or where land is completely disrupted by opencast mining, new parent materials are made available for the processes of soil formation to work upon. The changes which take place in these materials demonstrate how the early stages of soil formation take place. The direction or rate of soil formation may be changed gradually or abruptly when the other factors of soil formation change, for example climate, or where the natural processes are interrupted by the actions of human beings.

Time

Soils, like organisms, change with the passage of time and gradually develop new features as they proceed from youth, through maturity, to old age. Young soils retain many of the features of the parent material from which they developed, but as they become more mature, the pedological features increase at the expense of the characteristics of the parent material. This progression starts with the addition of the first fragment of organic matter to a new parent material and continues with the development, and increasing clarity, of the genetic horizons. By the time a soil is in equilibrium with its environment, it can be considered to be a mature soil. Most of the early forms of soil classification are based on the characters of the mature soil profile.

There are several examples of soil formation that has taken place over a known period of time. Often, these examples are related to catastrophic events, such as the eruption of Krakatoa in 1883, which was well-documented, or to the development of soils on lavaflows, landslide debris and earthflows. The retreat of glacier fronts has left behind an expanse of new parent material upon which soil formation has begun. Records of the earlier position of glaciers provide an opportunity to date accurately the beginning of soil formation.

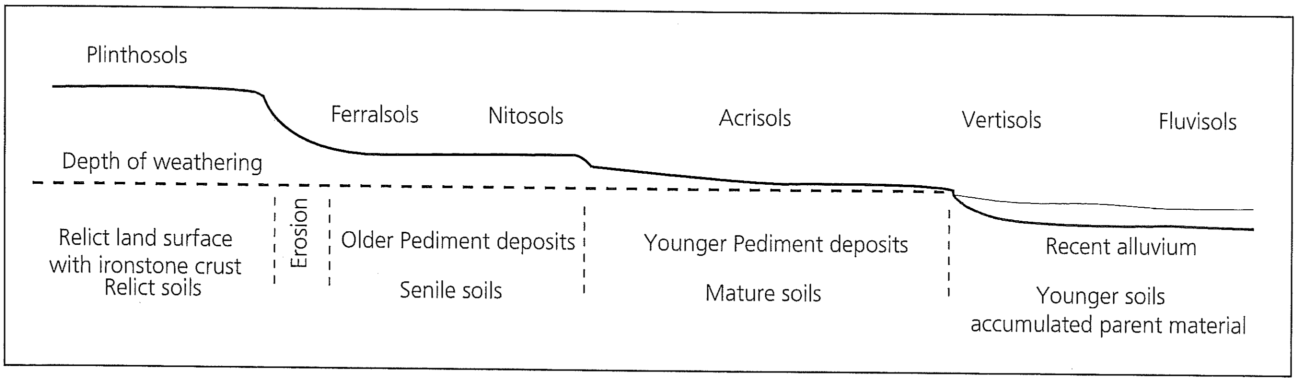
Examples are also available where the draining of a lake has exposed its floor to the processes of soil formation; one of the best-documented examples is the draining of the polders from the bed of the IJsselmeer in The Netherlands. The initial changes in the new parent material have been described as a process of **ripening**, during which profound physical, chemical and biological changes take place as the soil develops.

The development of a sequence of dunes parallel to the coast has occurred in several parts of the world, providing information on the rate of soil formation. Similar sequences of successive mining spoil heaps show changes taking place as soil development occurs, including accumulation of organic matter and acidification, which can be related to the time which has elapsed since the mining spoil heaps were dumped.

In many inter-tropical regions, it is possible to establish a sequence of soil development correlated with the morphology of the landscape, thus linking together two branches of geographical study. The position of a soil on the landscape gives its age relationships (Fig. 3.8). In the example, the time-scale of geomorphology and that of soil formation are working in parallel, although for most soils the time-scale of soil formation is shorter than that concerned with the formation of geomorphological features.

In the British Isles, it is generally accepted that soil formation dates from the end of the Pleistocene, about 10,000 years BP. Most of the soils formed during the interglacial periods were eroded by subsequent glaciation, but a few relict soils remain to give some indication of the earlier soil cover. Pockets of soil material, often eroded

Fig. 3.8 Age and position of soils on the landscape. Different geomorphic surfaces give rise to soils of different maturity.



and redeposited, that remain in Scotland and South Wales indicate warmer conditions during the last interglacial period. Evidence of soils formed in colder conditions is preserved within the sequence of glacial deposits in East Anglia.

Changes of climate are known to have occurred during post-glacial times and the effects of this have been preserved in soils. Following the Iron Age period of British pre-history, increased podzolization occurred in southern Britain. Many soil profiles have features which suggest that the present soil-forming factors are different from those of the past, reflecting climatic change. The evidence is to be seen in the lower soil horizons that have yet to be brought into equilibrium with the present environmental conditions.

Modelling soils

The modeller's approach, which aims to build in the mind's eye a simple but comprehensible picture of a complex reality in the real world, has five main components. The first of these concerns the **external variables** or factors such as temperature, precipitation and, in more recent times, the influence of pollutants on soils. Secondly, there are the **state variables**, the parameters of the soil system as shown by the amount of various constituents and concentrations of nutrients. Such chemical, physical and biological relationships may be linear or non-linear and can be demonstrated by mathematical equations. It is necessary to choose parameters carefully for specific soils. These parameters must be calibrated realistically by comparison with known observations. Associated with these parameters is the idea of **sensitivity**; for example the impact of some parameters may change dramatically with the seasons. It is necessary to **conceptualize** the processes in diagrammatic or mathematical form, demonstrating the interaction of the various elements of the model. Finally, the model has to be **verified** for its internal logic and the accuracy of its output. In recent years, models have been used extensively to obtain a better understanding of crop growth, nitrogen cycling, nutrient uptake, carbon content, organic matter turnover and soil–plant moisture relationships.

The functional relationship of the soil-forming factors shown by Jenny is but one extremely useful approach to the study of soils using a conceptual model. However, there are some real difficulties in applying this particular approach in every case. The significant problems of giving climate, organisms or parent material a value which can be substituted in the equation can only be resolved by the use of subjective decisions. Furthermore, it is almost impossible to isolate completely the variables: relief and climate are often closely related and with them the nature of the associated plant and animal ecosystems. Later, after attempting more complex soil–ecological equations, Jenny simplified his model to three state factors:

$$s = f'(Lo, Px, t)$$

where Lo = the initial state of the system, Px = the external flux potentials and t = the age of the system. Following this lead, Runge has proposed a factorial model which lays more stress on energy relationships:

$$s = f'(o,w,t)$$

in which o = organic matter production, w = water for leaching and t = time. Other modellers have attempted to use changes in chemical or mineralogical composition in models which try to balance the inputs and outputs of the soil system.

Since Jenny's *Factors of Soil Formation* was published in 1941, there have been many developments in the use of models. Two main types of model are used in soil studies: concrete models and conceptual models. The former implies experimental soil columns in the laboratory through which solutions can be percolated, as well as lysimeters in the field containing undisturbed soil profiles which can be weighed and the inputs and outputs measured. Conceptual models include mental, verbal, structural and mathematical forms.

An important step forward has been the recognition of the soil as an open system, stemming from the proposals of Simonson. The biogeochemical cycling of elements, as suggested in Chapter 1, can form a useful outline model for soil genesis, as well as illustrating the path of plant nutrients or pollutants through the soil and its associated plant and animal ecosystems (Fig. 3.9). Weathering of the rocks is an input which becomes the mineral part of

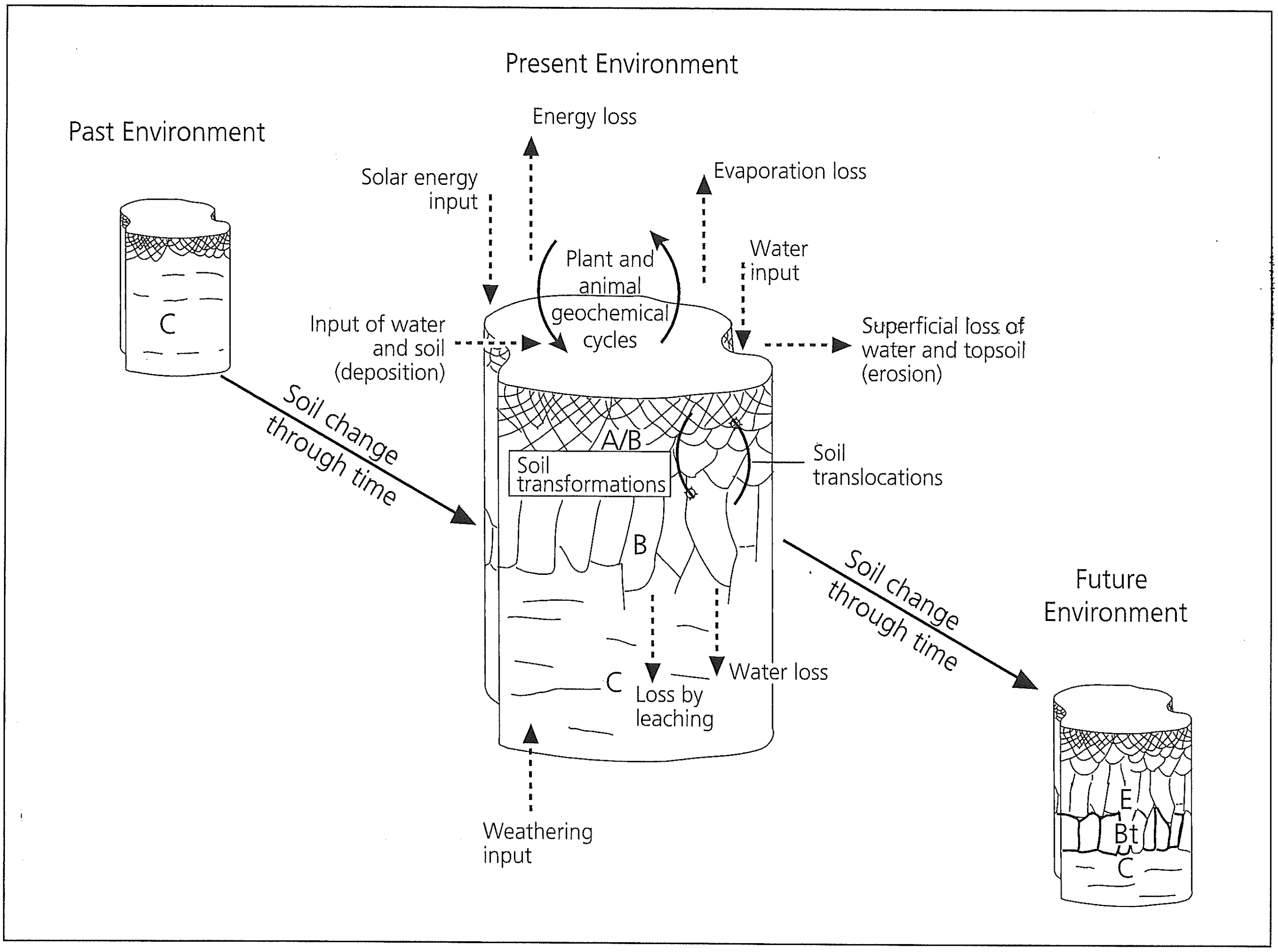


Fig. 3.9 The soil as an open system. A soil area, or pedon, is shown as part of a system evolving through time. Profile characteristics result from additions (inputs) and losses (outputs), as well as through translocations and transformations within the profile.

the soil; some of the elements released by weathering will be plant nutrients and these become an output of the soil system and an input of the associated biological cycle. Recycling will occur when the plant dies and its component elements are returned in the plant litter, decomposed and taken up again by a later generation of plants. Losses from the system occur when dissolved minerals are carried away by the drainage water, or crops are harvested.

An advantage of the systems approach lies in the idea of 'feedback' loops, which help to regulate and maintain the soil system. Any model of soils has to be a dynamic model because soils change over time. It may be a rapid change, as with changes of land-use, or it may be a slower change in response to the factors controlling their development. Some of the oldest tropical soils may well approach the steady-state or dynamic equilibrium advocated by systems theorists, but even these soils are subject to change if one of the factors of soil formation changes.

Modelling forms an important aspect of pedology when, with limited amounts of basic information, it becomes necessary to extrapolate from one set of data to another. A new area of study, which is evolving rapidly, involves **pedotransfer functions** to derive surrogate data for use when observations were not made at the time a soil was sampled. To date this has applied particularly to soil physical data: insufficient data were gathered in the earlier days of soil surveys, and are now expensive, time-consuming and labour-intensive to obtain. Most progress has been made with pedotransfer functions that estimate water contents at certain matric potentials, or estimate the soil-water retention or other physical parameters. The soil attributes used most commonly by modellers include particle size, organic carbon content and bulk density, and predicted values of soil-water retention, saturated and unsaturated soil conductivity have been used. These attributes of soils are necessary for modelling the movement of water, plant nutrients and pollutants through soils (transport models). They are incorporated in models of the uptake of nutrients by crops and are also used in models predicting soil erosion. With restricted funding for soil studies, the use of these pedotransfer functions becomes increasingly attractive, and they have the added advantage that they can be handled by computer and interfaced with geographical information systems to give a spatial dimension to the studies.

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4 Processes of soil formation

The processes of soil formation modify the regolith and give it the acquired characteristics which distinguish soil from parent material. These processes include weathering, clay formation, leaching, eluviation, podzolization, calcification, ferrallitization, salinization, alkalization, solodization, rubefaction, gleying, accumulation of organic matter, and pedoturbation. These processes are not mutually exclusive: podzolization and gleying can be seen to take place simultaneously in cool, humid regions of the world; gleying and salinization are frequently seen in low-lying areas of semi-arid parts of the world. The results of the soil-forming processes are to produce the characteristic features of soils. These are used in the identification and classification of soils.

Weathering

Originally, rock minerals were formed deep in the Earth's crust at temperatures and pressures very different from those at the surface. When eventually revealed at the Earth's surface by geological action, these rock minerals are unstable in their new environment and are susceptible to weathering by the processes of hydrolysis or acid decomposition, oxidation, reduction, hydration and carbonation.

Weathering is an important precursor of soil development but it also occurs alongside specific soil-forming processes, especially in shallow or young soils. It can be considered under the headings of either physical or chemical weathering. Physical weathering is most obvious in arctic and desert regions where free water is limited, but elsewhere chemical weathering is dominant in the production of soil parent materials. Water is critically important in the process of chemical weathering. Charged with dissolved carbon dioxide and the breakdown products of organic matter, rain-water forms an acid solution which attacks unweathered rocks and their constituent minerals. At the same time, the movement of water through the soil and regolith removes the soluble products from the site of weathering and so prevents equilibrium concentrations being reached in the soil solution. This ensures that rock decomposition continues.

Studies of the soil solution, the liquid component of the soil, have shown that weathering of the common feldspar minerals takes place in two stages: mineral breakdown followed by organic complexation. An example is the hydrolysis of orthoclase feldspar to kaolinite, releasing potassium ions and silicic acid into the soil solution. Feldspar minerals have a complex composition, reflected in their surface form, which is porous; this allows the acidic soil solution to penetrate and exploit any weaknesses present. Hydrolysis of clay minerals occurs because the dipole form of the water molecules orients them with their positively charged side towards the mineral. Their opposite, negatively charged side attracts hydrogen ions (i.e. protons, H^+) which, after satisfying the overall negative charge, enables other hydrogen ions to enter the lattice structure of the clay particle and exchange with potassium, sodium or calcium ions within the mineral, thus disrupting its structure. The process is encouraged further by the formation of organo-metal complexes (chelates), which enables the breakdown products to pass into solution and percolate down the soil profile.

Other weathering processes which take place in the regolith include oxidation and reduction. Iron and manganese minerals are particularly involved, and often their state of oxidation gives typical colours to soil horizons. Iron oxides and hydroxides are present as very small particles, coating other, larger mineral grains such as quartz. When the iron oxides are removed, the uncoated quartz grains give the typical grey colour of the albic E horizon. The chemical process of reduction also takes place and is very important in waterlogged conditions; reduction makes iron and manganese more soluble and oxidation–reduction processes produce the typical grey and yellowish-red colours of the mottling of gley soils. Dissolution removes many minerals, and carbonation affects calcareous material such as calcium carbonate, which is attacked by carbonic acid producing soluble calcium bicarbonate.

Weathering results in loss of the features of rock structure and leads to development of the regolith. In the temperate regions, depth of weathering is normally one or two metres only, but in the humid tropics much greater depths of weathered material can be seen. Extremely deep weathering profiles occur, in which the present-day soil is formed only in the immediate surface. Below the soil, in the parent rock, a series of different layers may be identified, separated by 'weathering fronts'. These layers include the ironstone, pallid and mottled zones of laterites, and a plasmic zone in which the fabric of the primary rock is destroyed and replaced by new soil parent material. At the base of the weathered profile lies the saprolite, where the rock is completely rotten but retains its rock structures, and the saprock, which still contains weatherable minerals. Some or all of these layers may be present, depending upon the erosional history of the site.

Clay formation

Soils developed from recent alluvial sediments may contain the minerals illite, vermiculite and chlorite, which have been formed elsewhere and transported to their present site. However, in many soils, as a result of the chemical elements liberated during the weathering process, the ingredients are present for new clay minerals to be formed within the soil. Some clay minerals are formed simply by alteration, but clay formation, together with the production of different colours by oxidation, and structural changes caused by wetting and drying cycles, is an important contributor to the development of a **cambic B horizon** (Plate A7).

In semi-arid or sub-humid regions, active weathering on the interfluve areas releases silica and magnesium which become concentrated in alluvial lowlands, where recombination may occur to produce smectite clays. Where the soil solution becomes over-saturated with aluminium hydroxides and silica, allophane together with the minerals imogolite and halloysite contribute to the resultant clay fraction. This occurs particularly in regions where volcanic ejecta are weathering to form soil parent materials.

In the humid tropical regions, kaolinite, the iron oxides ferrihydrite and goethite, and the aluminium oxide gibbsite form a group of **low activity clays**. These clays, with a very low cation exchange capacity (CEC) and a low base saturation, are produced throughout the inter-tropical region, where the silica content of the soil solution is very low and the concentration of other elements is also very dilute.

Leaching

This is the term given to the process by which soluble constituents are removed from the soil. In any part of the world where rainfall exceeds evaporation, readily soluble salts are dissolved by water percolating downwards. Consequently, soluble salts are removed in the drainage water and do not persist in the soil profile. Calcium carbonate and even sparingly soluble minerals are dissolved and carried away over a long period of time (Fig. 4.1).

In most humid climates, however, not all of the calcium, magnesium, potassium and sodium is present as free salts in the soil; some is held as cations by the clay–humus complex. A process of proton exchange takes place in which hydrogen ions are exchanged for the cations, gradually acidifying the soil. After prolonged attack by acid weathering and leaching, as has occurred in some soils of the humid tropical regions, only quartz, kaolinite, gibbsite and a few other very stable minerals remain in the soil, in part protected by a coating of iron oxides composed of goethite, with a yellowish-brown colour, or hematite with a bright red colour. In association with clay formation, leaching is responsible for the development of the cambic B horizon.

The leaching process is checked when soils are limed for agricultural purposes. Plants also tend to reduce the effect of leaching, as they bring elements in solution from the subsoil through their roots to the stems and leaves. These elements are then released by the death and decomposition of the

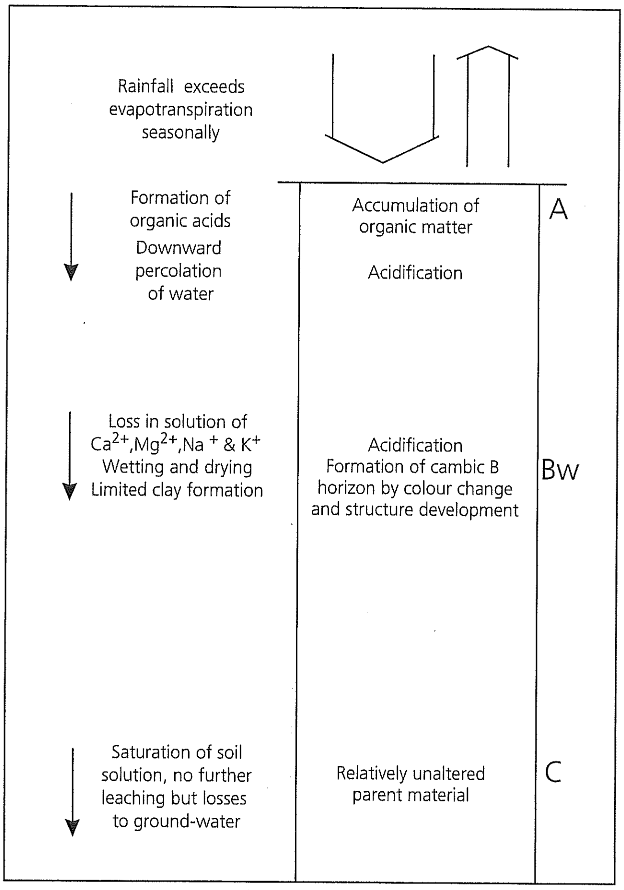


Fig. 4.1 The process of leaching.

plant on the soil surface. Burrowing animals, too, tend to bring material to the surface by their activities, and this is also in opposition to the leaching process.

Clay eluviation

In the past, the term 'eluviation' simply implied the removal of substances from the A horizon of a soil. Recently, it has become customary to distinguish between loss by solution (leaching) and eluviation, which refers specifically to the loss in suspension of material from a soil horizon. Finely dispersed humus and clay particles, as well as other weathering products, can move as colloidal suspensions from upper, **eluvial** horizons to **illuvial** horizons lower in the profile, where they are redeposited (Fig. 4.2). This process appears to be encouraged by a climate in which a period of desiccation results in the soil shrinking and cracking. As the soil dries, the suspended material is deposited from the soil solution on the sides of the peds and along pores.

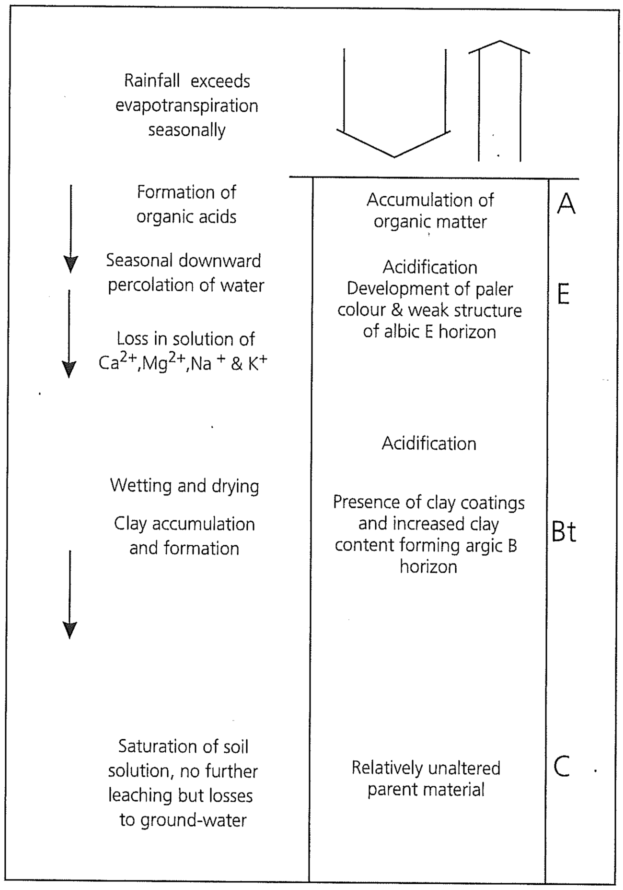


Fig. 4.2 The process of clay eluviation.

Pedologically, one of the most important results of eluviation is the development of a B horizon enriched in clay, which is referred to as an **argic B horizon** (also called an argillic horizon, a Bt horizon or a 'textural' B horizon) (Plate A8). In some soils the clay content is appreciably raised by this phenomenon of eluviation. The results of eluviation can be seen as clay coatings on subsurface soil structures, which are also observable by microscopic examination in thin sections of soil samples.

The argic horizon often has a prismatic or subangular blocky structure. In order to qualify as argic, a horizon must have an increased clay content: if the overlying horizon has less than 15 per cent clay, the argic horizon must have over 3 per cent more; if the overlying horizon has between 15 and 40 per cent clay, the argic horizon must have 20 per cent more; and if the overlying soil has a clay content of more than 40 per cent then an argic horizon must have 8 per cent more.

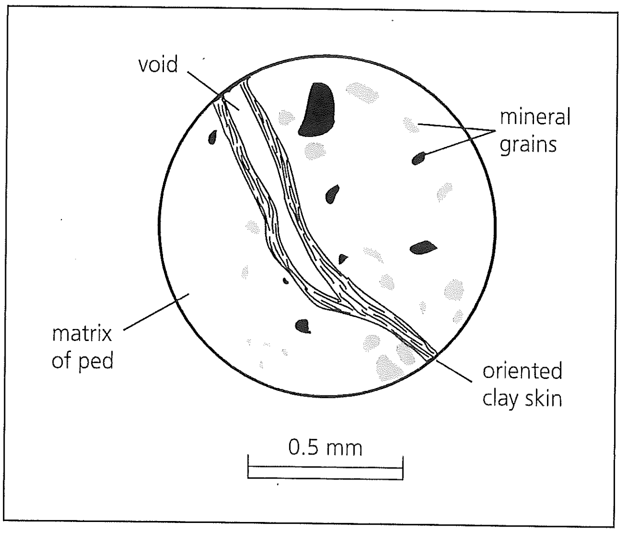


Fig. 4.3 A thin section of soil as seen through a microscope, showing oriented clay lining the walls of a pore.

Eluviation is regarded as a purely mechanical washing of fine particles suspended in the soil solution from the upper into the lower horizons. For this reason the process is also referred to by the French name of *lessivage* and the group of related soils in France are called *Sols Lessivés*. In well-developed examples, clay skins can be seen with a hand lens or even with the naked eye (Fig. 4.3). It is also possible to observe the downward movement of silt particles from the eluvial horizon of grey wooded soils in Canada and the Derno-podzolic Soils of Russia. This results in a characteristic white tonguing of the E horizon into the B horizon, a feature which is referred to by the term 'glossic'.

Podzolization

The process of podzolization is prevalent in soils of the cool humid parts of the world, but it is also responsible for the development of extremely deep soils on quartzitic sands in tropical regions. The results of podzolization are readily distinguishable because the processes that operate are more severe and the profile formed is more distinctive, both in appearance and in its physical and chemical properties, than in the case of soils formed only by leaching or eluviation (Fig. 4.4).

Podzolization involves the development of an extremely acid humus form known as mor, which

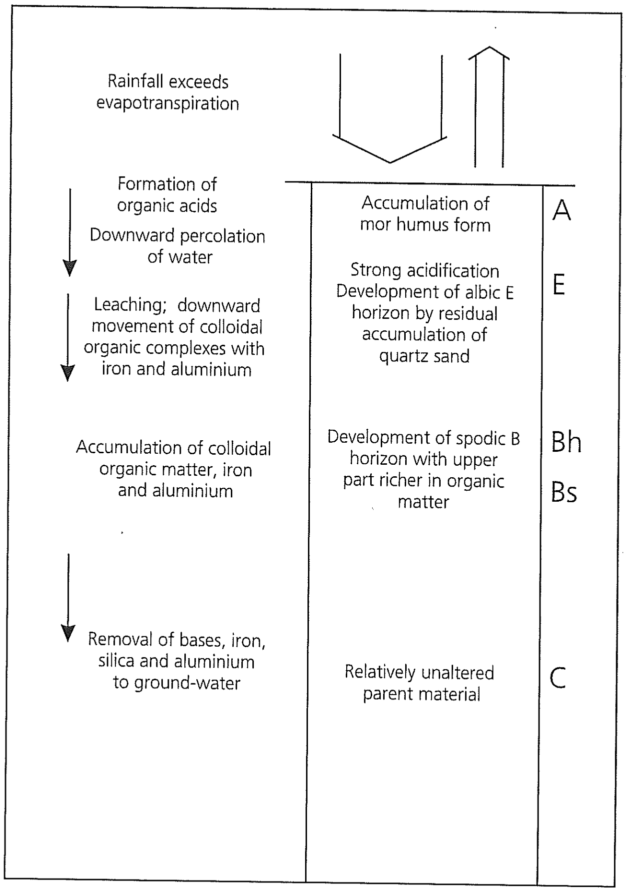


Fig. 4.4 The process of podzolization.

forms because the rate of decomposition of debris from plants such as heath or coniferous trees is slow, allowing extremely acid litter, fermentation and humus horizons to accumulate (Fig. 2.8). Rain-water falling on the vegetation acquires soluble organic breakdown substances from the plants; as it percolates through the surface organic horizons, further organic breakdown substances are added to it before it enters the mineral soil. This acidified soil solution is capable of disrupting the structure of the clay minerals, releasing the constituent elements. Silica and aluminium from the clays, iron from iron minerals and coatings on mineral grains, complexed with organic matter, are mobilized and removed from the surface horizons as the solution percolates downwards through the soil profile.

As even the most sparingly soluble elements are eventually removed by this process, there is a tendency for the almost insoluble quartz grains of sand to form a relative accumulation in the immediate subsurface as the other soil components are removed. In this way, the strongly bleached, grey **albic E horizon**, typical of Podzols, gradually forms (Plate A6). The A horizon, with its accumulation of organic matter, remains thin in the absence of earthworms to mix the organic and mineral materials.

The iron and aluminium oxides, which were mobilized, together with the organic matter, from the eluvial horizons and moved down the soil profile in the soil solution, are eventually deposited in an illuvial B horizon which has been given the name **spodic B horizon** (Plate A10). This is defined as occurring below an A horizon or an E horizon and at a depth greater than 12.5 cm. A spodic B horizon must have one or more of the following: a sub-horizon more than 2.5 cm thick, continuously cemented by a combination of organic matter with iron or aluminium or with both; a sandy or coarse-loamy texture with distinct dark pellets of coarse silt size or with sand grains covered with cracked coatings of organic matter and aluminium with or without iron; and one or more sub-horizons must have the following features:

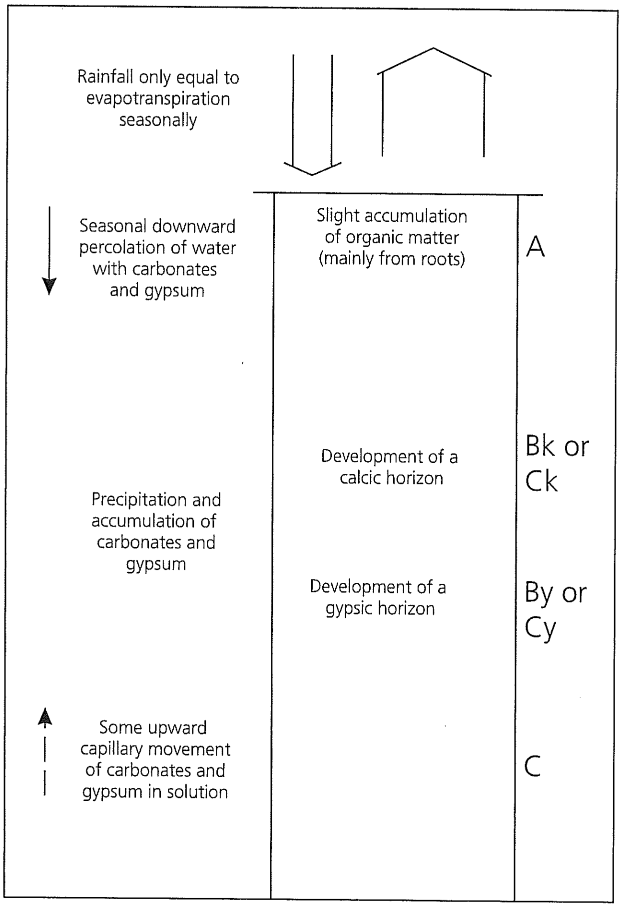
(a) if there is 0.1 per cent or more extractable iron, the ratio of iron plus aluminium extracted by pyrophosphate at pH 10 to the percentage of clay is 0.2 or more, or if there is less than 0.1 per cent extractable iron, the ratio of aluminium plus organic carbon to clay is 0.2 or more; and
(b) the sum of pyrophosphate-extractable iron plus aluminium is half or more of the sum of dithionite-citrate extractable iron plus aluminium; and
(c) the thickness is such that the index of accumulation of amorphous material in the sub-horizons that meet the preceding requirements is 65 or more. This index is calculated by subtracting half the clay percentage from the CEC (expressed in $cmol_c kg^{-1}$ clay) at pH 8.2 and multiplying the remainder by the thickness of the sub-horizon (in centimetres); the results of all the sub-horizons are then added.

Strongly acid conditions developed in all Podzol soils limit the range of soil fauna present as well as the range of plants that can be successfully grown. Earthworms are noticeably absent and the soil undergoes little faunal pedoturbation, so the pronounced horizons of the Podzol profile are allowed to develop.

Calcification

The process of calcification is characteristic of soils in low-rainfall semi-arid and arid areas and continental interior situations. Leaching is slight, and although downward movement does take place, the soluble constituents are not removed from the soil profile. After seasonal rains or snowmelt, these soils are wetted only to a depth of 1 to 1.5 m, where the impetus of the downward-percolating moisture is lost and it begins to be drawn back to the surface to re-evaporate (Fig. 4.5). Successive wetting and drying cycles in the soil lead to the deposition of calcium carbonate in a **calcic horizon** (Plate A12), usually in the lower B or upper C horizons of the soil profile. To qualify as a calcic

Fig. 4.5 The process of calcification/gypsification.



horizon there must be a zone of at least 20 cm where calcium carbonate is accumulating and the amount must be in excess of 15 per cent of the fine earth and 5 per cent more than in a lower horizon.

Calcium carbonate accumulation may be in the form of fine particles diffused throughout the soil matrix (Plate A31), as concretions of soft, powdery lime (Plate A26), or as discrete nodules. However, in some cases it is in the form of continuous, strongly cemented layers which may occur at depth or at the surface where erosion has exposed them. The name **petrocalcic horizons** has been given to these layers (Plate A13).

In arid areas, processes similar to those which lead to the accumulation of carbonates in the soil profile can result in the accumulation of gypsum in a **gypsic horizon** (Plate A14). The gypsum can occur as scattered crystals through the matrix of the soil (Plate A32), as crystallaria ('desert roses') or as a massive **petrogypsic horizon** (Plate A15). Where both calcic and gypsic horizons occur in the same profile, the gypsic usually lies below the calcic horizon.

Ferrallitization

The process of ferrallitization is characteristic of soil formation in the humid tropical regions of the world. In the past, this process has been referred to as laterization, latosolization, kaolinization or desilicification; these terms have become confused in their definition and use, so the term ferrallitization is preferred. In simple terms, this process involves the net loss of silica, the formation of kaolinite and the relative accumulation of the sesquioxides of iron and aluminium (Fig. 4.6). The accumulation of Fe and Al gave this process of soil formation its name. On basic parent rocks the process is fairly rapid, leading to the formation and accumulation of the iron minerals goethite and hematite, but gibbsite may be formed instead of kaolinite. On acid rocks the process is much slower and clay formation is restricted to the kaolinite group of clay minerals. In both cases, the clay minerals frequently are coated and cemented with iron oxides. The resulting soil material is usually porous and therefore freely drained, and is yellowish-brown or red in colour, depending upon the

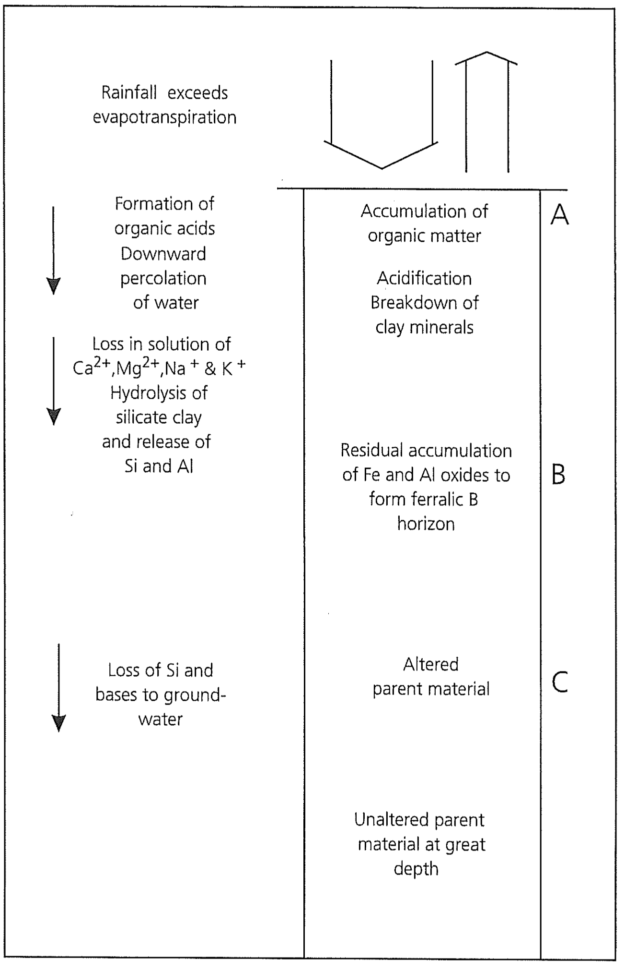


Fig. 4.6 The process of ferrallitization.

degree of hydration of the iron oxides. Because of the cohesion given by the iron oxide coatings the clay does not easily disperse in water and so the soil is relatively resistant to erosion.

The process of ferrallitization is accompanied by strong leaching of the soil, so pH values are low. The rapid decay and recycling of the elements contained in leaf-fall from the forest trees keep the limited amounts of plant nutrients and bases in circulation between plant and soil. Geomorphological stability and a prolonged exposure to the humid tropical environment allow time for the production of a highly weathered, low base-status **ferralic B horizon** (Plate A11), alternatively known as the oxic horizon.

To qualify as a ferralic B horizon ferrallitization must have proceeded to such a degree that the soil material is more than 30 cm thick, has a particle size of sandy loam or finer, contains less than 10 per cent weatherable minerals, possesses less than 5 per cent rock structure, and any clay increase must be gradual. The ferralic B horizon is also characterized by a CEC equal to or less than 16 $cmol_c kg^{-1}$ clay as measured by the ammonium acetate method at pH 7.0. In extreme cases the amount of exchangeable bases (Ca, Mg, K, Na) may be reduced to less than 1.5 $cmol_c kg^{-1}$, with the presence of unbuffered 1 M KCl-extractable Al or a delta pH (pH KCl - pH H_2O) of +0.1 or more. In these cases the soil is said to possess **geric properties**.

Salinization

In arid climates the rainfall is irregular and is insufficient to remove soluble salts from the soil. With occasional rain they are moved down the profile, but when dry conditions resume they may be drawn upwards again. In semi-arid areas salts may be washed from soils of the upland areas, and there is a redistribution of salts into the soils on the lower parts of the landscape. In most cases, the occurrence of soils affected by salinization is associated with those areas in arid and semi-arid regions where there is a high ground-water level and imperfect or poor natural drainage. Frequently, these areas are the alluvial plains and other areas which in moister climates would produce considerably greater yields of crops.

Regrettably, many soils are suffering salinization as a result of excessive use of irrigation water without adequate drainage, which has raised the ground-water to a level where capillary rise brings salts into the soil. The water is evaporated, leaving the salt in the soil. These soils often develop a surface encrustation of salt and are known as Solonchak soils (Fig. 4.7). Such soils are described as having **salic properties** (Plate A16). This term refers to an electrical conductivity of the saturation extract of more than 15 dSm^{-1} at 25°C within 30 cm of the surface, or of more than 4 dSm^{-1} if the pH exceeds 8.5. The salt may originate from a salt-rich geological substratum, or it may be derived from salt sea-spray, blown inland by onshore winds, which gradually accumulates in the unleached soils of arid and semi-arid areas.

Soils in which the presence of salt in the profile results from natural processes are called primary saline soils. In hot, dry countries the use of irrigation water containing even small quantities of soluble salts can result in the salinization of soils as the water is evaporated. The resulting soils are described as secondary saline soils as they only occur through human interference in the natural environment. In both primary and secondary saline soils, the depth to the water table is of critical importance.

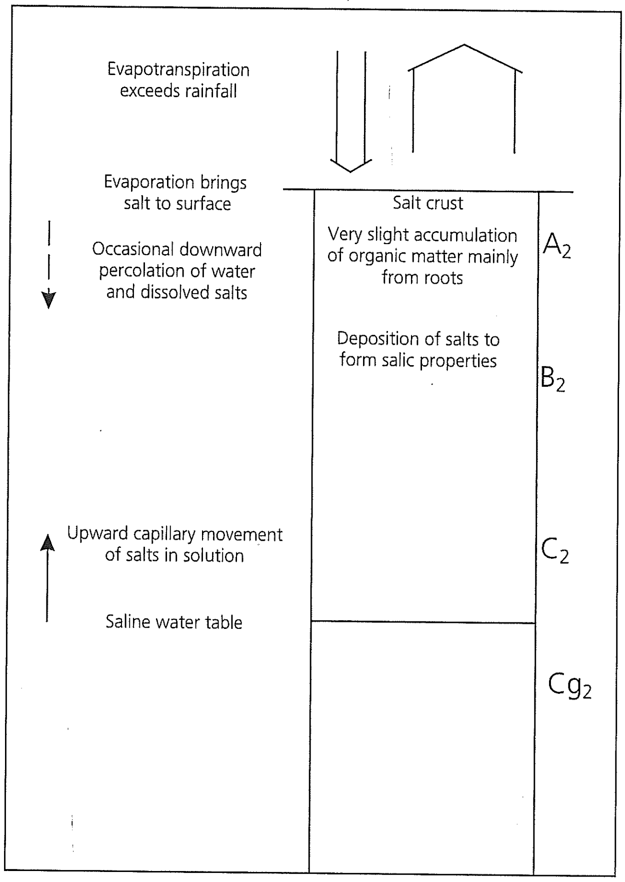


Fig. 4.7 The process of salinization.

Alkalization

The process of alkalization occurs when sodium ions come to dominate the exchange positions of the clay–humus complex of a soil. This is achieved when slight leaching removes the soluble neutral salts. The solubility of calcium and magnesium is lower than that of sodium, which therefore remains in the soil solution after the divalent ions have been precipitated. Drying can also concentrate sodium ions remaining in solution so that they monopolize the exchange positions on the clay–humus complex, giving the soil a **natric B horizon** (Plate A9). A natric B horizon is defined as one which has the properties of an argic horizon, but with a columnar structure and with more than 15 per cent of the exchange capacity occupied by sodium ions within the upper 40 cm of the horizon. The soils become strongly alkaline (pH >8.5) and sodium carbonate is formed within the soil. The process results in Solonetz soils; these are extremely difficult to manage as the organic-rich topsoil becomes completely dispersed when wet, and the natric horizon itself is either very plastic when wet or as solid as concrete when dry.

Solodization

The process of solodization, or removal of sodium ions from the clay–humus complex, results in a range of soils from the Solodized Solonetz to the Solod. In these soils, the metal cations, mainly sodium, are gradually leached out and once the neutral salts have been removed the exchangeable sodium hydrolyses, increasing the hydroxyl ion concentration in the soil solution. This results in dispersion of the organic and mineral colloids. As the metal cations of the salts are gradually replaced by hydrogen ions, a strongly acid soil eventually results, called a Solod. The process of change from a salty soil to a Solod can be initiated by a change in climate, such that increased rainfall affects soils, by water from snowmelt accumulating in low-lying areas to give increased leaching, or by irrigation.

Rubefaction

Certain soils, formed under cool wet winters and hot dry summers, have a marked red colour caused by an even distribution of iron oxides throughout the profile. The term was introduced to describe the reddening observed in many soils of the Mediterranean countries of Europe. An irreversible loss of water from iron oxide gels leads to the production of a hematite coating on soil particles, which gives the bright red colour.

The reddening process increases with the age of the soil and so there is some correlation between colour and soil history: many brightly coloured Red Earths in Australia clearly have had a long pedogenetic history. Outside the present-day areas of Mediterranean climate, bright red palaeosol remnants along the coast of south Wales in the UK have been used as evidence for a warmer climate during the last major interglacial period of the Pleistocene.

Gleying

The presence of water in the soil for long periods brings about anaerobic conditions, as explained in Chapter 2. This can happen in all climatic zones, and the process of gleying produces features which enable the pedologist to recognize the hydromorphic character of the soils and to consider these soils together as a group of Gley or Hydromorphic Soils. Poorly drained soils frequently occur on plateau sites or on the lower parts of the landscape, and are often developed from parent materials related to those of adjacent freely drained soils.

The process of gleying involves the reduction of iron compounds and either their complete removal from the profile or their segregation into mottles or concretions. Usually the gleying process is accomplished with saturated conditions in the presence of organic matter. The metabolic activity of the bacteria is responsible for the oxygen deficit, and the reducing conditions develop best where the soil solution is stagnant. A strongly gleyed soil or gleyed horizon is frequently an unrelieved grey or bluish colour, but where some oxygenation takes place there will be mottles of rust-coloured ferric oxide.

For soils developed in a strongly seasonal climate, a number of these reactions are combined into a process of soil formation referred to as **ferrolysis**. Cycles of oxidation–reduction in the soil environment allow the leaching of displaced cations in the reduced phase and the acid weathering of clay minerals in the oxidized phase at the beginning of each cycle. The process is driven by energy derived from bacterial decomposition of soil organic matter. In the reduced state, iron is changed to the ferrous form and replaces adsorbed basic cations, which are partly leached out of the soil. As the soil is oxidized again in the dry season, the exchangeable iron is changed from the ferrous to the ferric state with the production of protons (H^+), which contribute to the next cycle of disruption of clay minerals.

Two basic types of Gley Soil are widely recognized by pedologists. Where water is held upon a slowly permeable or an impervious horizon within 50 cm of the soil surface, producing mottling, a Surface-water Gley or Pseudogley Soil results, with **stagnic properties** (Plate A18); where water rises up through the soil from an impermeable layer at depth beneath the soil, to give mottling within 100 cm of the soil surface, a Ground-water Gley or True Gley is produced, with **gleyic properties** (Plate A17). It is usual to find Ground-water Gley Soils in areas which receive an inflow of surface- and ground-water. In this case, the soil must be permeable if the water is to be able to move between the soil structures and saturate the lower soil horizons of the profile.

With anaerobic conditions in the soil, the decomposition of plant residues on the surface is slow and their incorporation into the soil is minimal. Therefore peat often develops on soils with a high level of ground-water. The colour of the soil in these profiles is usually an unrelieved grey, reflecting the presence of ferrous iron compounds. If the soil dries occasionally there may be a few mottles of rust-coloured ferric oxide.

Surface-water Gley Soils can occur almost anywhere on the landscape where a slowly permeable horizon, such as a Bt horizon or an iron pan, prevents the free downward passage of water through the soil profile. Localized gleying results, particularly on the surfaces of pores and peds, which become grey. In contrast, the ped interiors retain the brighter colours of ferric iron, giving an overall impression of mottling when revealed in a soil pit or exposure. Soils with such mottles scattered throughout their profiles closely resemble the Pseudogley Soils described by German authors.

Accumulation of organic matter

Plant leaves and other debris that fall onto the soil surface, if undisturbed, may accumulate to form an organic horizon that is not saturated with water for more than a few days in the year. This horizon is composed of three layers. In the litter layer (L), the process of decomposition (already begun on the plants) turns the easily recognizable leaves a dark brown colour, but the type of leaf is still recognizable. In the fermentation layer (F), plant litter loses its distinguishing characteristics, and eventually in the lowest layer (H) becomes a black amorphous substance referred to as **humus**. Humus is a complex polyphenolic organic substance with many active chemical functional groups. These unincorporated organic layers, lying on the surface of the mineral soil, are not normally saturated, and are collectively referred to as the O horizon. They are not, however, a diagnostic horizon for classification purposes.

Decomposition is a complex process during which a succession of soil fauna break the material down into smaller pieces, with fungi and bacteria eventually causing the complete disintegration of the plant tissue into chemical elements and compounds which can be recycled by the plants through their roots. The dark-coloured, more resistant humus remains to darken the soil surface horizon. Although an average figure for the turnover of all organic matter is given in the literature (20–22 years), some organic matter fractions are fairly rapidly recycled, with a turnover time of between two and three years. Use of radioactive carbon dating enables estimation of the mean residence time of the stabilized and more resistant material. The turnover time of these more stable forms of soil carbon, measured at Rothamsted, England, ranges from 1450 years in the topsoil to 12,100 years at 2 m below the surface. Analysis of the average of ^{14}C age in some representative topsoil samples of soil organic matter from North America ranges from 210–440 $\pm$ 120 years in surface soils in Iowa and North Dakota, to 25,300 years $\pm$ 9 per cent for a frozen peat soil in Alaska.

The process of darkening of soil material by organic substances is known as **melanization**. The presence of organic substances in the soil confers stability to the soil structures. Organic substances also act as a source of energy for the soil fauna, and the gradual release of elements from the humus is a source of plant nutrients.

Accumulation of incorporated organic material in soils takes place in four distinctive diagnostic surface horizons. These are the ochric, mollic and umbric A horizons, and the histic H horizon.

Most of the soil organic matter comes from above-ground leaves and stems of plants. Where the accumulation of humus is relatively slow, because of the nature of the organic debris available or the prevailing climatic conditions, incorporation is slight and an **ochric A horizon** is formed (Plate A4). An ochric A horizon is a thin A horizon which has too light a colour, too high a chroma and too little organic carbon to be mollic or umbric (see below). An ochric A horizon can be hard and massive when dry, and may be of the moder type in humid areas or mull in drier climates.

The type of humus associated with calcareous soils is mull, produced by the natural vegetation of grasses and the action of earthworms. Grasses have well-developed root systems which, when alive, ramify through the soil, and when dead supply large amounts of organic matter to the soil. The aerial parts of the grasses also return to the soil as organic matter when they die back after the growing season, or are burnt. The lack of leaching in semi-arid climates and a combination of summer drought and winter frost in continental areas, typically the steppes or prairies, limit the rate of organic decomposition so that over many years a very rich and deep **mollic A horizon** (Plate A2) accumulates.

Essential features of the mollic horizon include the following: even after ploughing to a depth of 18 cm, the surface horizon should have a strong structure which is not massive or hard when dry; it should have colours with a chroma of less than 3.5 when moist and a value darker than 3.5 when moist and 5.5 when dry, and a colour value that is at least one unit darker than the C horizon; the base saturation should be 50 per cent or more; the organic carbon content should be at least 0.6 per cent, although if there is much finely divided lime in the horizon the colour criteria may be waived and then the organic carbon content should be more than 2.5 per cent; thickness must be more than 10 cm if resting directly upon hard rock or a petrocalcic, petrogypsic horizon or a duripan; where the solum is less than 75 cm a mollic horizon must be more than 18 cm, but where soil depth is more than 75 cm it must be 25 cm (this includes transitional horizons in which A horizon characteristics are dominant); finally, there should be less than 250 $mg\,kg^{-1}$ citric-soluble phosphate present. Similar dark-coloured, strongly structured horizons that have all the preceding characteristics except that the base saturation is below 50 per cent are termed **umbric A horizons** (Plate A3).

In conditions of very poor drainage, peat formation is encouraged in association with the gley soils present. Basically, there are two types of peat: the moor peat which is strongly acid, and the fen peat which is neutral or mildly alkaline. Further subdivision can be made based on botanical composition, structure and degree of decomposition of the organic remains. The characteristics of these two peat forms result from the way in which they have developed. The acid moor peat develops on upland areas where high rainfall results in the leaching of all the bases so that acidity dominates and there is very slow decomposition of plant debris, forming a **histic H horizon** (Plate A1). As a consequence, the organic matter accumulates above the level of the mineral soil and can eventually develop into a raised bog if the rainfall is sufficient. A raised bog is one in which the plants are sustained from the nutrients brought in by the rain.

Fen peats are developed where organic matter accumulates in waters liberally supplied with bases and which have a neutral or mildly alkaline pH. These conditions occur mainly in lowland alluvial situations. When adequately drained, the soils developed upon such parent materials are usually very fertile with high land values, as in the Fens of eastern England.

Pedoturbation

All clays swell and contract to some degree when they pass through cycles of wetting and drying, and this results in the development of structure and cambic horizons. Pedoturbation occurs where the movement of soil material is more obvious. Where the supply of bases is plentiful, as in the seasonally dry tropics, low-lying areas of the landscape frequently develop soils rich in montmorillonite. This clay mineral has a great capacity to expand as it takes up and releases moisture; as it does so, the individual peds expand, and press and move against each other causing polished faces to be formed, called slickensides. The surface expression of this pedoturbation is frequently a mound-and-hollow micro-relief, often referred to as 'gilgai' after an

Australian aboriginal word. Many soils throughout the world have these features, referred to as **vertic properties** (Plates A22 and A23).

In sub-arctic regions, freeze–thaw cycles can also result in the mixing of different layers of soil material in the subsoil, forming features described as 'festoons'. In many soils, mixing by soil fauna is an important feature of their development. The lack of well-developed soil horizons in Brown Earths is partly caused by the continual mixing of the soil by earthworms in the process of faunal pedoturbation.

Conclusion

The processes of soil formation may be described as 'bundles' of processes which operate within a soil. These processes, which may vary in intensity, can be seen more simply as one or more of the categories (additions, losses, transfers or transformations) from the generalized theory of soil genesis proposed in 1959 by Simonson. These four categories include the following.

Additions

- water, from rainfall, irrigation, seepage or capillary rise from ground-water
- radiant energy from the sun
- fresh and decayed organic matter added to the surface (O horizon) or incorporated in the surface horizon (Ah horizon)
- organic matter added to the Bh horizon of Podzols
- iron compounds added to the Bs horizon of Podzols and the Bg horizon of Gley Soils
- silicate clay added to the Bt horizon as clay coatings or void infillings
- calcium and magnesium carbonates added to A, B or C horizons
- silica or other cements added to indurated horizons

Losses

- moisture through evaporation, transpiration or drainage
- heat
- organic matter from the A horizon
- silicate clay from the A and E horizons
- carbonates from the A, B or C horizons
- silica by weathering of primary minerals
- iron through leaching and gleying
- aluminium through leaching

Transfers

- movement of mineral grains from open to tighter packing arrangements (pedocompaction)
- mixing of organic and mineral soil components by wetting and drying, freezing and thawing and animal activity (pedoturbation)
- movement of iron compounds into nodules and concretions as a result of gleying

Transformations

- decay of plant remains into humus
- weathering of primary minerals to form silicate clays, oxides and hydroxides of iron and aluminium
- transformation of soil components into new forms (e.g. clay into clay coatings).

Further reading

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5 Soil classification

Classification is only a contrivance to order a subject, perhaps to understand and remember its content more easily, and it should certainly show how any one part of the subject is related to the whole. A classification also serves to create a 'language' or nomenclature for exchanging knowledge, and is indispensable for conceiving mapping units on a soil map. As far as soils are concerned, classification is still in the process of development, and throughout the last century, numerous attempts were made to group soils into natural, homogeneous classes with common properties. A final valuable role of classification is that it serves as a basis from which further enquiry can proceed.

Soils are not discrete individuals, like animals or plants, but gradually change laterally from one to another – they are said to form a **continuum** upon the Earth's surface. Because soils are not readily separated from one another, classification has proved to be a considerable challenge, and the history of soil science is marked by the various attempts to bring order into the subject.

Although earlier forms of soil classification are now mainly of historical interest, they remain relevant to present-day pedology because they show how concepts developed, and many of the names introduced in earlier classifications remain in the literature and may still be used, even though the system of classification itself has fallen into disuse.

Soil classification is fundamental to a study of the soils of the world, and so it is an important consideration in a book devoted to world soils. Equally, classification is essential for making a soil map. However, classification requires a wide knowledge of the subject before it can be fully understood. Students new to pedology may find it helpful to read further and gain some insight into the broad major types of soil formation before returning again to this chapter. For the time being, you are recommended to proceed with the approach adopted at the end of this chapter in which the soils of the world can be traced using the key in Fig. 5.1.

Origins

The earliest attempts to classify soils may be traced to China 4000 years ago when nine classes were recognized, but the evidence suggests that this was more a classification of land than of soil. In Europe, the earliest documented assessment of soils dates from Greco-Roman times, when soils were placed in categories ranging from best to worst for the growth of crops.

Throughout much of historical time, soils were grouped by their texture: clayey soils, loamy soils, sandy soils and organic soils. Although simple to use and with very practical significance for the agriculturalist, this simple means of classifying soils includes such a wide range of different soil types that, outside a purely local setting, texture alone cannot be used meaningfully in a classification system unless combined with other criteria.

In Chapter 1 it was stated that soil studies in modern times developed from two branches of learning: geology and agro-chemistry. No early forms of classification emerged as a result of the agro-chemical approach, but in the 19th century the growth of geological knowledge led to a recognition of the importance of the underlying rocks as parent materials for soils. Somewhat inevitably, this linkage was transferred to soil classification, so names such as 'limestone soils', 'sandstone soils' and 'alluvial soils' came into common usage amongst those who cultivated the land. These terms can still be heard in use, even though perfectly good pedological names are available.

The modern period

It is generally accepted that present-day soil science