SOILS
THEIR ORIGIN, CONSTITUTION, AND CLASSIFICATION
SOILS
THEIR ORIGIN, CONSTITUTION AND CLASSIFICATION
AN INTRODUCTION TO PEDOLOGY
BY
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CURTIS FLETCHER
MARBUT

Curtis Fletcher Marbut was born in Lawrence County, Missouri, on July 19th, 1863. He died, on a scientific mission, at Harbin, Manchuria, on August 25th, 1935. The Soil Survey of the United States, under his direction from 1910 to 1935, became the model for such enterprises throughout the world. The loftiness of his scientific ideals, the nobility of his character, and the kindliness of his disposition, won the admiration and affection of all who knew him, and an influence that reached beyond the bounds of his own country. Although he had passed the ordinary span of life, his death was untimely and his loss irreparable.
"Habet enim rationem cum terra, quae numquam recusat imperium nec umquam sine usura reddit quod accept, sed alias minore, plerunque maiore cum faenore. Quamquam me quidem non fructus modo sed etiam ipsius terrae vis ac natura delectat."

CICERO : De Senectute.
PREFACE TO THE THIRD EDITION

TWELVE years have passed since the last revision of this book and, although soil research has been hampered by war conditions, advances in our knowledge have been sufficiently important to call for many changes from the former edition. I have not been able to embody, or even to mention, all the contributions made to the subject during these years. Some selection was inevitable, and I have naturally drawn on those resources of new information that seemed most likely to help in my own particular approach to the subject. One reason for economy in the use of new work is the desirability of keeping "Soils" to a convenient size. My original aim was to exhibit the soil in perspective, and I hope that in the present edition the subject—so much greater and more complicated than it once appeared—may still be thus seen. Further developments may make it difficult to give a documented account in one moderately-sized volume. A later edition might need to be separated into two volumes, one on the origin, constitution, and properties of soils, and a second on soil systematics and geography, but the end of such a development might be a "Handbuch," which is beyond my present intention.

I hope that "Soils" may still, as I believe it has done in the past, introduce the beginner to a new and fascinating subject and tell workers in other branches of science something of this young study. Perhaps it may also pose a few critical questions for the more advanced student. Although my aim has been to give readers "a right judgement in all things," I know that much of what I write will be out of date in twenty years. But the unceasing search for truth will go on.
In addition to the acknowledgments of help in earlier editions, very warm thanks are due to those colleagues who have so kindly helped me in this revision. I am again particularly grateful to Dr. R. K. Schofield for his valued counsel in the chapters dealing with the physics and chemistry of soil. I am indebted also to Professor G. Aubert (France and French Africa), Dr. C. R. Van der Merwe (South Africa), Mr. C. G. Stephens (Australia) and Dr. L. I. Grange (New Zealand) for supplying material about their respective countries for inclusion in the chapter on Soil Geography. Finally, I am indebted to Dr. C. E. Kellogg for so generously allowing me to use certain diagrams and photographs. He has thus continued the kindly interest in my work shown by his predecessor, Dr. C. F. Marbut, whose memory is again perpetuated in the dedication of this book.

G. W. ROBINSON.

University College,
Bangor,
PREFACE TO THE SECOND EDITION

In presenting the second edition of "Soils," I feel that some indication should be given of the changes which have been made from the original edition, and of the principles which I have followed in making such changes.

In the first place, I have made many small variations from the first edition by correcting erroneous or misleading statements and by clarifying obscure explanations. It is, perhaps, too much to hope that those parts of the first edition which survive have been completely purged of faults. Secondly, I have added some material which, although not modifying the original account, may augment its "corroborative detail." Here it has been necessary to exercise a certain economy in view of the abundance of material available. Thirdly, I have re-cast those portions of the book which appeared to require amendment in view of recent advances. These changes are mainly in the chapters on the Pedogenic Processes, on the Clay Complex, on Base Exchange, on Soil Moisture, and on Soil Classification. Fourthly, I have added additional examples in the chapters on the Soil Groups and have somewhat amplified the chapter on Soil Geography. Lastly, I have withdrawn the appendix on Methods of Analysis in view of the publication of Mr. C. H. Wright's book on this subject. The narrative has been increased by 63 pages. When allowance is made for excisions and replacements, it will be seen that the amount of fresh material is considerable. About 150 new bibliographical references have been added.

The three years which have elapsed since the appearance of the original edition have been marked by great activity in soil investigation. Not only have large numbers
of papers been published; but, what is still more important, new ideas have emerged and earlier ideas have been modified or clarified. I have not found it possible to embody even the majority of these investigations in the new edition. Indeed, it has not been my object to record everything that has been done in this branch of study. "Soils," as originally presented, was my view of the subject and was not intended to be an impersonal and impartial account. The references to the literature, given therein, were those which I deemed necessary to document that account. And so, in the second edition, I have only included such new work as has modified my earlier views, and have by no means attempted to record everything that has been written on soils in the intervening years. I venture to hope that I may escape the charge of undue economy of truth.

A book which proposes to give a view of the whole of a subject should be more than a mere compilation of facts. It should have, indeed, something of the character of a work of art and interpret the ideas of the author. It should be a picture rather than a photograph. And so, the facts which have come to my notice have fallen into a certain perspective and have produced my picture of the soil. Another pedologist might arrange the same facts in a different perspective and produce another picture. The reader who views these different pictures may be counselled to adopt the attitude of the great English Prime Minister who, when confronted with apparently contradictory pronouncements by his colleagues, "marvelled not so much at the diversity of testimony as at the many-sidedness of truth."

I hope I may secure the indulgence of those of my colleagues who have laboured with distinction at particular aspects of pedology and who may not find my treatment of those aspects completely satisfying. Thus, although my treatment of soil moisture may be inadequate for the soil physicist, I venture to hope that it may prove not wholly useless and misleading to the soil micro-biologist. Similarly, the soil micro-biologist may excuse the shortcomings of my
account of his subject if I have succeeded in enlightening the soil physicist on this aspect of pedology.

I wish to express my thanks for the entirely helpful criticisms of the original edition which I have received from colleagues in all parts of the world and trust that I may still have the benefit of their counsels. Further, I am greatly indebted to Dr. R. Kenworthy Schofield for the use of diagrams on pF relationships and for his helpful suggestions for the chapter on Soil Moisture; to Dr. E. W. Russell for the use of his diagram on crumb formation; to Mr. W. C. Davies, Curator of the Museum of the Cawthron Institute, New Zealand, for the picture of the Pakihi profile; to Mr. W. Morley Davies for his picture of a podsol profile; and to Dr. C. E. Kellogg, Chief of the United States Soil Survey, for several illustrations which add materially to the interest of the book.

G. W. ROBINSON.

UNIVERSITY COLLEGE,
BANGOR,
Christmas, 1935.
PREFACE TO THE FIRST EDITION

The student of Don Quixote may remember the criticism of Cervantes on one of his own works: "He propounds something and concludes nothing." On looking over my proofs, I am aware of the aptness of this judgement to the present work. My aim has been to give, within the compass of a book of moderate size, a general view of pedology—the study of soils. Recent developments, however, have so increased the scope of the subject that only the broadest treatment is possible if an adequate perspective is to be given within the limits proposed.

The first section of the book is occupied with the origin, constitution, and properties of soils, and, since the standpoint adopted is the exhibition of soils in their natural relationships, I have exercised some selection by dealing mainly with those topics which appear necessary for this purpose. The next section is devoted to the description, with illustrative examples, of the chief soil groups of the world. This is followed by a discussion of the problem of classification, and an account, given with due reserve, of the geographical distribution of soils. In the remainder of the book, I have dealt with soil surveys and soil analysis, and have concluded with a brief discussion of the inter-relationships of soils, plant growth, and agriculture.

I have written primarily for those who are interested in the soil as an object of study in itself, and secondarily for those whose interest lies in its economic or geographical significance; but it may be hoped that the book will prove of value to botanists because of the importance of the soil in
ecology, and to geologists because of the part played by the soil in the weathering cycle.

Since the subject is here treated in its purely philosophical aspect, I have avoided the consideration of certain subjects, such as lime requirements and exchange acidity, which, though occupying important places in the literature, are definitely problems of applied pedology. I have also refrained from the discussion of those enquiries, belonging rather to physical chemistry or physics than to pedology, in which the soil appears merely as a research material. The omission of these subjects implies no adverse reflection on their value or on the ingenuity of the investigations by which they have been illustrated: it is imposed by the necessity of giving, within the available space, the fullest possible picture of soils in their natural relationships.

In developing my account, I have referred only to such investigations as appear necessary to illustrate the main arguments. The bibliography is not, therefore, as full as might be desired by the research worker. The references given at the end of each chapter will, however, lead the reader to the sources of further information. The frequency with which I have used examples from the work of my own laboratory will not, I trust, be held to betoken an undue estimate of its importance. I have made free use of this work because it is most familiar and lies most readily to hand for purposes of illustration.

My cordial thanks are due to those authors who have so kindly allowed me to make use of material from their published works or private communications. Special acknowledgments are also due to Messrs. Paul Parey, Berlin; Gebr. Borntraeger, Berlin; The Macmillan Company, New York; The Williams and Wilkins Company, Baltimore (Soil Science); and The Cambridge University Press (Journal of Agricultural Science) for permission to reproduce tables or diagrams from their publications.
I am particularly indebted to Mr. G. H. Freeman (Messrs. Thos. Murby & Co.), whose sympathetic co-operation has made the author's share of the work of publication a pleasant and interesting task.

Finally, I wish to record my indebtedness to the colleagues, past and present, who have worked with me at Bangor. They have helped in many tangible ways, but most of all by their loyal support throughout many happy years of study.

G. W. Robinson.

University College,
Bangor,
Easter, 1932
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CHAPTER I

INTRODUCTORY

The study of the soil, from its close connexion with the important art of agriculture, has occupied, until recently, an intermediate position between the pure and the applied sciences. Most of the earlier investigations on soils were directed towards the solution of problems arising from practical agriculture, and this circumstance has introduced complications both in the statement of the subject and also in the methods of research employed. The investigator of the relationships between the growing plant and its environment is thus in some danger of failing to secure an adequate view either of the soil itself or of the subject of plant nutrition.

Students of the soil are, therefore, under a debt to such leaders as Dokuchaiev and Glinka in Russia, Ramann in Germany, and Hilgard in America, who succeeded in creating a body of effort directed towards the investigation of the soil as a definite branch of natural philosophy. For this study, the name pedology* has been proposed. In spite of the reluctance in some quarters to accept this reasonable name for the study of soils, it can hardly be maintained that there is a satisfactory alternative.

Pedology is the study of the soil from the standpoint and by the methods of pure science. In one aspect, it is a branch of geology, since it is concerned with the superficial strata of the regolith†, in which most of the important

*Pédology, from πέδος = ground or earth. It should not be confused with pédology (or pedology), a term sometimes used in America for the study of children.
†Regolith (or better, rhegolith), the fragmental unconsolidated débris, mantling the rocks of the earth’s crust.
processes of weathering and denudation take place. In another aspect it is the study of the physical and chemical properties of a complicated colloidal system. In a third aspect it is the study of a complex flora and fauna in relationship to soil environment. These three aspects are closely interdependent, for a complete knowledge of the origins and metamorphoses of soils involves an understanding of soil constitution, which itself is closely related to soil microbiology.

The adoption of the view that pedology should be regarded as an independent branch of study and treated as a pure science does not imply any depreciation of the practical significance of the subject. On the contrary, it may be submitted that the study of the genesis and constitution of soils and the development of a philosophical system of classification will yield, as corollaries, practical results that could hardly be obtained by direct search. The manner in which pure science has contributed to practical life during the past century affords ample justification for this view of the relative importance of the pure and the applied standpoints in the study of natural phenomena. As the discoveries of the pure chemist and the pure physicist have found applications in daily life, so we may expect that, when the constitution of the soil is more perfectly understood, applications to practical agriculture will also be forthcoming.

The most important advance in pedology during recent years has been the recognition of the soil profile as the unit of study. To an earlier generation of investigators, the individual soil meant a sample of soil, generally from the superficial layer occupied by the roots of plants. We now know that an adequate conception of the soil can be obtained only from a study of all the horizons or soil strata down to the parent material. And whilst for some purposes it is sufficient to study the properties of samples taken from the individual horizons of a profile, a complete picture cannot be obtained until the laboratory studies are supplemented by observations on natural profiles in the field,
The investigations of recent years have been for the most part directed towards the elucidation of the soil viewed as a static system. Much of this work belongs as much to colloid chemistry and physics as to pedology. But, eventually, the pedologist must go into the field and study his material under natural conditions. The past years have been fruitful in perfecting the instruments of research. In the writer's opinion, the centre of interest must now shift more and more from the laboratory to the field. It is therefore to be expected that the future contributions to pedology will tend, in an increasing measure, to be descriptions of new types of soil, and studies of the chemical, physical, and biological processes that take place under field conditions.

The soil profile—the succession of horizons down to the parent material—gives expression to all the processes of soil formation, including physical and chemical weathering, and the vertical translocation of soil constituents by water movements. The character of these processes, and hence the character of the soil profile, is mainly governed by climate. Climate also controls the type of natural vegetation, which in its turn reacts on profile development. But the parent geological material also plays a part, and under certain conditions may prove the most important factor in differentiating soils from each other.

Fully developed soil profiles are not everywhere in evidence. In some cases the soil profile is immature, as, for example, in soils developed on recent alluvial or aeolian deposits, where time has not sufficed for the full expression of the pedogenic processes. In other cases, the profile may have been truncated by erosion, whereby some or all of the horizons have been removed, the present-day vegetation being established on a lower horizon of the original profile. The natural profile may also be disturbed by cultivation, which results in a confounding of all horizons within the reach of the implements of tillage.

Since the character of the soil profile is affected by the mode of action of the pedogenic processes and by the
nature of the parent material on which these processes have operated, it will be readily understood that the possibility of variety in the final result is very great. Yet we shall see in the sequel that the distribution of the major soil groups is mainly determined by climate, whilst geology is chiefly of significance in transition regions and in the later stages of classification. Topography also enters as a factor and may profoundly modify the operation of the regional climatic influences.

The horizons lying below the soil profile are regarded as being the parent material from which the soil profile has been differentiated. This is not always the case, for in superficial deposits of mixed origin the soil profile may have developed from material different from that lying immediately below. In other cases more than one parent material may be present in vertical succession within the profile itself. For many purposes it is convenient to regard pedogenesis as consisting of two groups of processes, namely (1) the weathering processes that give rise to the parent material, and (2) soil profile development from the parent material formed by weathering. Whilst these processes are ideally regarded as successive, they may proceed simultaneously. The former case is illustrated by the development of a soil profile from deep drifts or other unconsolidated deposits, and the latter case by the development of shallow profiles in decomposing rock.

In a relation of the origin, constitution, and classification of soils, the familiarity of the subject might almost be held to render a definition superfluous. Yet, viewing the soil as the most superficial layer of the regolith, the student is at once faced with the difficulty of deciding where the study of the soil ends and where the domain of the geologist begins.

Where a thin layer of soil lies directly on the rock from which it has originated, there is every justification for regarding all material down to the unaltered rock as within the province of the pedologist. The position is less clear
when we consider soils derived from deep unconsolidated deposits such as alluvium, loess, or glacial drift. To restrict the study of the soil to the upper layers in which plant roots ordinarily fulfil their activities would be to exclude horizons which clearly fall within the sphere of action of the pedogenic (soil-forming) processes. To define the soil as those horizons affected by the pedogenic processes would be, of course, to state an identity. Nevertheless, a study of deep sections suggests that below a certain depth, generally not more than a few feet, the strata are purely geological, and only indirectly affected by the pedogenic processes which operate in the upper horizons. On the whole it seems preferable, in the present state of our subject, to decline a definition and merely to indicate that the range of interest of the soil student is not confined to the depth utilized by plant roots and subjected to tillage by the cultivator. By adopting this position, the pedologist claims the right to study the pedogenic processes in their widest significance. The domain of pedology can only be defined by the natural limits of enquiry and may come to engross a considerable portion of dynamic geology. Indeed, studies of contemporary pedogenic processes may be expected to throw light on many of the problems of sedimentary petrology.

Considered in quantitative relationship to the lithosphere, the regolith is of trifling thickness, whilst the soil, the superficial stratum of the regolith, is even more insignificant. Yet the importance of the soil in dynamic geology can hardly be over-estimated, for it is the seat of the most active operation of epigene weathering and is also that part of the earth’s crust immediately subject to denudation. The soil is thus an essential phase in the weathering cycle, and its study is a branch of geology in its broadest sense.

Considering soil as material, its essential ingredients are (a) mineral matter, which has originated from rock by the action of a series of weathering processes, (b) organic matter, which has originated from the residues of natural vegetation
and organic manures, (c) soil moisture, containing substances in colloidal or in true solution, and (d) soil air. The mineral matter includes particles of varying grade from stones and gravel down to sub-microscopic particles of clay. In it may be recognized particles of rock-forming minerals, and material which is either amorphous or else consists of minerals whose crystalline character can be revealed only by X-ray analysis. The latter material is called the colloidal clay, or clay fraction, and confers on soils and clays those properties, such as plasticity and cohesiveness, which distinguish them from mechanically comminuted rock. Whilst the physical disintegration of rocks can account for the recognizable rock and mineral fragments in soils, the clay fraction results from processes of chemical weathering, mainly silicate hydrolysis. We shall, in a later chapter, discuss this process with more minute attention.

The organic matter of soils is present mainly in the form of dark-coloured amorphous material, which we may designate as humified organic matter or, simply, humus. It has originated naturally from the residues of vegetation, but in cultivated soils organic manures have also contributed. Fragments of undecomposed plant-remains, leaves, stems, roots, and seeds, can also be recognized, but from 60 to 90% of the organic matter of soils is in the humified state. Organic matter added to the soil may be either oxidized to carbon dioxide or, under conditions which we shall discuss later, humified. The amount of organic matter in a soil is thus governed by the extent to which it is enriched by plant residues, and also by the relative intensity of the oxidative and humifying processes. Whilst desert soils may contain fractional percentages of organic matter, the prevalence of the humifying processes in humid climates may result in the production of peat soils consisting almost entirely of organic matter. In ordinary British soils the proportion of organic matter varies from 3 to 10%.

Although for many purposes it is convenient to study the colloidal clay and humus separately, in the soil itself they
occur in intimate admixture or even in loose chemical combination as the so-called clay-humus complex. Some of the most important properties of the soil are governed by the amount and character of this complex.

A soil is not completely defined by a quantitative specification of its constituents. It possesses a definite structure, which, however, is subject to change even in the same individual soil. Structure is of the highest importance both as an aid to diagnosis in soil classification and also in connexion with the problems of plant growth in soils. Its study is at present being actively pursued.

In writing on the subject of soils, some difficulty is experienced in securing a satisfactory order of presentation. Thus, it is difficult to discuss the pedogenic processes without reference to the colloidal complex, and this, in its turn, involves the study of organic matter and of base exchange. The inter-relationships of different branches of the subject have sometimes entailed a certain amount of repetition and the appearance of the same topic at more than one point in the narrative. The reader who is making his first acquaintance with the subject may therefore find it advantageous to re-read some of the earlier chapters in the light of later chapters.

The chapter on Soil Classification has been placed after the chapters in which the principal soil groups are described. This seemed advisable because the problems involved cannot be appreciated until the reader has gained some knowledge of the material which it is proposed to classify.

GENERAL BIBLIOGRAPHY

The following books may be recommended to those who wish to study the subject in greater detail:


E. Ramann. Bodenkunde. Julius Springer, Berlin. Also a classical work, it forms a bridge between Hilgard and contemporary authors.


The following journals are devoted wholly or partly to publications on the study of soils:


Soil Science. The Williams and Wilkins Co., Baltimore, U.S.A.


CHAPTER II
GENERAL VIEW OF THE CONSTITUTION OF THE SOIL

PROXIMATE CONSTITUENTS

The most superficial consideration of soils will reveal the immense variety of their composition and structure; but there are, nevertheless, certain groups of constituents present in all soils, and this imparts a unity to their study. Briefly, these groups are (1) mineral matter; (2) organic matter; (3) soil moisture; and (4) soil air. The last two are, strictly speaking, dependent variables and, in amount and character, controlled by the mineral and organic matter. We may thus conceive of the soil as a three-phase system with an indefinite number of components.

The solid phase is composed of the mineral matter and the organic matter. The liquid phase is in equilibrium both with the solid phase—the mineral and organic matter—and the gaseous phase—the soil air. But the equilibrium is continually shifting owing to the variations in temperature and in water content, the drain of nutrients by plant roots, and the activities of micro-organisms. With these variations are associated corresponding changes in the composition of the gaseous phase—the soil air. It is obvious, then, that the solid phase of the soil offers a surer basis of study than the rapidly changing and dependent liquid and gaseous phases. Indeed, for many purposes, the study of the soil is synonymous with the study of soil mineral matter and soil organic matter in the dried laboratory sample. But, when it is necessary to link up the study of the soil with that of
the growing plant, a more complete knowledge of the conditions in the soil solution and the soil air becomes essential.

Considering, then, soil as it appears in the laboratory, we note that it is a mixture of (a) mineral material, which consists of the products of physical and chemical weathering of rocks, and (b) organic matter, which consists, similarly, of the more or less decomposed residues of plant materials and, to a less extent, of animal remains and excreta.

It is not possible by any known means to effect an exact separation of the mineral and organic material of soil, and this constitutes a serious obstacle to its study. By boiling with hydrogen peroxide (15-30%), however, practically the whole of the organic matter is either oxidized to carbon dioxide and water or else decomposed to form soluble products that can be removed by filtration. It is thus possible to obtain the mineral portion of the soil practically without admixture. Unfortunately, no method is known, or seems likely to be discovered, whereby the organic matter can be isolated unaltered, and it is thus necessary to study it either in predominantly organic soils or else in the presence of a great excess of mineral matter.

The mineral matter of soil consists of particles varying in size from stones and gravel down to the submicroscopic particles of colloidal clay. Where a complete range is present, it will be seen that the larger particles consist of rock fragments, the intermediate particles of rock-forming minerals, and the smallest particles of the secondary products of chemical weathering. Very approximately, we may say that, in ordinary European soils, fragments above 2 mm. in diameter are mainly rock fragments; particles from 2 mm. to 0.002 mm. in diameter are mainly original mineral fragments, whilst the material below 0.002 mm. in diameter consists principally of the secondary products of chemical weathering.

In certain tropical soils, material of secondary origin may occur among the coarser fractions. Indeed, in highly
THE CONSTITUTION OF THE SOIL

developed laterites the surface layer may consist almost entirely of coarse material of secondary origin containing comparatively small amounts of original minerals.

MECHANICAL COMPOSITION OF SOILS

The mineral portion of soil consists of particles varying in size from stones to clay. A general knowledge of the mechanical composition of the soil, i.e., the relative proportions of the different sized particles, is of the highest importance both for the purpose of characterizing the soil and also because it can afford an indication of its physical properties, more particularly in respect of its behaviour under cultivation. Mechanical analysis, the series of operations whereby mechanical composition is determined, is the most important laboratory examination to which soil is submitted.

Practical farmers recognize a soil property commonly known as texture. When a farmer describes a soil as a clay, a heavy loam, or a light loam, he is describing its texture. Experienced soil surveyors can distinguish perhaps 15 to 20 grades of texture. As there is, among related soils, a correlation between texture and mechanical composition, there is a tendency to regard mechanical analysis as a means of accurately specifying texture. It should be realized, however, that mechanical composition is not a direct measure of texture. It gives an estimate of texture only as a result of observed correlations between mechanical composition and texture.

It is not our purpose, in the present chapter, to enter into the details of mechanical analysis as an analytical procedure; yet some allusion to the laboratory aspect of this subject is necessary in order to understand the significance of the results obtained.

In the first place, it is not generally possible to include the larger particles, gravel and stones, in the statement of mechanical analysis. Their inclusion would introduce diffi-
culties both in sampling and also in the actual laboratory procedure. It is therefore customary to make an arbitrary separation of the coarser particles by sieving. Although not universally adopted, the usual convention is to reject all material retained by a 2 mm. round-hole sieve. If desired, the proportion of this coarser material can be determined and expressed separately. The mechanical analysis, then, refers to material whose particles are all less than 2 mm. in diameter. As stated above, the material of coarser grade consists principally of unweathered rock fragments.

Two questions at once confront the analyst. In the first place, it is necessary to decide upon the number of fractions to be distinguished and to define these fractions by particle-size. Much confusion formerly existed on this point owing to the use of different scales in different countries. It has been decided by the International Society of Soil Science that the following fractions shall be distinguished:—

<table>
<thead>
<tr>
<th>Fractions</th>
<th>Diameter limits (mm.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coarse Sand</td>
<td>2.0-0.2</td>
</tr>
<tr>
<td>Fine Sand</td>
<td>0.2-0.02</td>
</tr>
<tr>
<td>Silt</td>
<td>0.02-0.002</td>
</tr>
<tr>
<td>Clay</td>
<td>&lt;0.002</td>
</tr>
</tbody>
</table>

In addition to its simplicity, the above scale, the so-called International scale, has the advantage that the successive particle-size limits are at equal logarithmic intervals.

Secondly, it is necessary to decide as to the basis on which the results of mechanical analysis shall be expressed. Here again there is a choice of procedure. The results may be expressed either on the air-dry soil, which contains a few per cent. of hygroscopic moisture, or on the soil dried at 100-105°C. There is also a variety of practice with regard to the inclusion of organic matter and calcium carbonate in the fractions, and as to whether the figures given shall refer
to ignited or oven-dry weights. In the method now adopted internationally, calcium carbonate and humus are removed by preliminary treatment, and the results refer to the carbonate-free mineral material, dried at 100-105°C. A mechanical analysis thus shows the percentages of the four fractions, and of the calcium carbonate, if present; the difference from 100 is principally the organic matter, of which, however, a separate determination must be made, since it is here loaded with all the errors of the mechanical analysis. It should be added that in the method at present used, whilst organic matter is to a large extent removed in the preliminary treatment, the removal is not complete, so that actually the fractions are slightly magnified by the inclusion of such organic matter as survives the treatment used for this purpose, namely, digestion with 6% hydrogen peroxide. The error thus introduced is relatively unimportant for ordinary purposes.

The validity of a mechanical analysis depends on the thoroughness with which the soil is resolved into its constituent particles by the dispersive treatment that precedes the actual mechanical analysis. For this reason a considerable amount of work has been directed towards discovering the most efficient method of preliminary treatment. The aim of this preliminary treatment is to break down compound aggregates and to bring the soil to its prime particle structure. Many of the earlier analyses were untrustworthy because complete dispersion was not effected, with the result that a proportion of the material analysed consisted of compound aggregates and not of ultimate particles.

The method of dispersion now adopted by the International Society of Soil Science1*, known as the International A-method, consists in boiling with 6% hydrogen peroxide to destroy humus, treatment with 0.2 N-hydrochloric acid and washing to remove carbonates and bases combined with clay and humus, and final dispersion by shaking in 0.008 N-sodium hydroxide.

*For numbered references see Bibliography at end of chapter.
Since the organic matter in soil does not generally exceed 10%, the mechanical analysis will usually account for 90% or more of the soil unless notable proportions of calcium carbonate are present. Dispersion having been effected by preliminary treatment, the coarsest fraction, namely, that consisting of particles between 0.2 mm. and 2.0 mm. in diameter is separated by means of sieving, using a 70 mesh brass wire-cloth sieve, whose apertures are slightly less* than 0.2 mm. square.

The finer fractions are determined by the application of a principle enunciated in Stokes' Law, namely, that the limiting velocity of a particle falling in a fluid is proportional to the square of its diameter. The separations can be effected by repeated decantation of the soil suspension, choosing appropriate settling times and depths, e.g., 8 hours and 10 cm. for clay. After each decantation of the supernatant suspension, water is again added; the suspension is thoroughly stirred, again adjusted to the mark, and allowed to settle again. The process of stirring up, settling, and decantation is repeated until, at the end of the time chosen, no more material remains in suspension. The material removed by decantation may be collected and weighed, or it may be estimated by the loss in weight of the material from which it has been removed.

Another method used in mechanical analysis depends on a determination of the concentration of a settling column of soil suspension after a given time at a given depth. For example, to determine the clay, a sample of suspension may be taken at 10 cm. from the surface after settling for 8 hours. The sampling is performed by means of a pipette introduced into the suspension, and the concentration found, expressed as a percentage of the original concentration, gives the percentage of clay, i.e., of material having a settling-velocity of less than 10 cm. in 8 hours.

*The error thus introduced is inappreciable with new sieves. With wear, the apertures become enlarged. See B. A. Keen and W. B. Haines, J. Agric. Sci., 1923, 13, 467-482.
The principles underlying mechanical analysis are discussed here in order to show that the lower particle-sizes used in mechanical analysis are purely inferential. Apart from the coarsest fraction, each fraction is defined in terms of a settling-velocity; the initial assumption, based on the work of A. Atterberg, being that a particle-diameter of 0.002 mm. corresponds with a settling-velocity of 10 cm. in 8 hours at 20°C. In the writer’s opinion, the use of particle-diameter is open to criticism, and may be misleading. Only if it were possible to make the separations throughout by sieving would the use of particle-size be justifiable. On the other hand, whilst definition of the fine particle-sizes in terms of settling-velocity is more logical, the fact still remains that the coarsest fraction can only be separated by sieving. It is, however, possible, by application of Stokes’ Law, to define the coarser particles in terms of settling-velocity or its logarithm. The correspondence of the two methods of definition is shown in the following table:

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Diameter Limits</th>
<th>Settling-Velocity (V)</th>
<th>Log V</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coarse Sand</td>
<td>2.0-0.2 mm.</td>
<td>347.6 cm./sec.</td>
<td>2.5412</td>
</tr>
<tr>
<td>Fine Sand</td>
<td>0.2-0.02</td>
<td>3.476</td>
<td>0.5412</td>
</tr>
<tr>
<td>Silt</td>
<td>0.02-0.002</td>
<td>0.03476</td>
<td>2.5412</td>
</tr>
<tr>
<td>Clay</td>
<td>&lt;0.002</td>
<td>0.0003476</td>
<td>4.5412</td>
</tr>
</tbody>
</table>

It will be seen that, whether the fractions are expressed as diameter limits or velocity limits, assumptions are unavoidable; but, in view of the importance of the finer fractions, and the frequent use of the principle of subsidence to determine still finer fractions, the use of settling-velocity or its logarithm appears to the writer to afford a more satisfactory basis.

In Table I are given examples of the mechanical analyses of some typical soils.
<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sandy Loam Shropshire</th>
<th>Light Loam Anglesey</th>
<th>Heavy Loam Anglesey</th>
<th>Rendzina clay Czechoslovakia</th>
<th>Clay East Africa</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coarse Sand</td>
<td>66.6</td>
<td>27.1</td>
<td>13.6</td>
<td>0.9</td>
<td>1.0</td>
</tr>
<tr>
<td>Fine Sand</td>
<td>17.8</td>
<td>30.3</td>
<td>17.4</td>
<td>7.1</td>
<td>4.1</td>
</tr>
<tr>
<td>Silt</td>
<td>5.6</td>
<td>20.2</td>
<td>24.7</td>
<td>21.4</td>
<td>7.9</td>
</tr>
<tr>
<td>Clay</td>
<td>8.5</td>
<td>19.3</td>
<td>35.1</td>
<td>65.8</td>
<td>82.8</td>
</tr>
<tr>
<td>CaCO₃</td>
<td>nil</td>
<td>nil</td>
<td>nil</td>
<td>2.0</td>
<td>nil</td>
</tr>
<tr>
<td>Org. Matter, etc.</td>
<td>dissolved in pre-treatment</td>
<td>1.5</td>
<td>3.1</td>
<td>9.2</td>
<td>2.8</td>
</tr>
<tr>
<td></td>
<td>Total ...</td>
<td>100</td>
<td>100</td>
<td>100</td>
<td>100</td>
</tr>
</tbody>
</table>


| Dissolved Sesquioxides | 0.4 | 1.5 | 1.4 | 0.7 | 0.4 |

**MECHANICAL COMPOSITION AS A CONTINUOUS FUNCTION OF PARTICLE-SIZE.**

The expression of the mechanical composition of a soil as percentages of a limited number of fractions is open to objection on account of the necessity for making arbitrary particle-size or settling-velocity divisions. Such divisions could only be justified, apart from their convenience, if it could be shown that each limit corresponds with a marked and discontinuous change in properties. For example, the separation of silt from fine sand at 0.02 mm. diameter would be completely justified if it could be proved that there is an essential difference in constitution and properties between particles of 0.021 mm. and particles of 0.019 mm. diameter. Such a proof cannot be furnished, for very little difference exists. The separation of clay at 0.002 mm. has more justification. At or about this particle-size there is a marked change in constitution and properties. Whilst the material
above 0.002 mm. consists mainly of rock fragments and chemically unweathered mineral particles, the material below this size consists mainly of the colloidal products of chemical weathering and is the chemically reactive portion of the soil. Whether 0.002 mm. is the most satisfactory limit remains to be ascertained. There is some reason for thinking that a more satisfactory separation might be obtained by the choice of a lower limit, say, 0.0006 mm., corresponding with log \( V = 5.5 \). It should always be remembered, however, that the actual basis of separation is the settling-velocity.

On account of the obvious drawbacks to arbitrary size limits, considerable effort has been devoted to the perfecting of methods whereby the mechanical composition of soils can be expressed as a continuous function of particle-size. S. Odén⁴ first attempted this by means of an apparatus in which the mechanical analysis of a soil or clay is automatically recorded by measuring the accumulation of falling sediment on a balance pan. G. Wiegner⁵ also devised an apparatus in which the change in density of a column of sedimenting suspension can be continuously recorded. E. G. Richardson⁶ proposed a nephelometric method in which the changes in concentration of a settling column are recorded photo-electrically.

**MECHANICAL COMPOSITION CURVES**

G. W. Robinson⁷, using the pipette method of analysis, depending on the change in concentration of a settling suspension with time, has shown that, if the mechanical composition of a soil or clay be expressed by means of a curve connecting summation percentages with the logarithm of settling-velocity, smooth curves are obtained which exhibit certain regularities. The significance of such curves will be appreciated by reference to *Figs. 1 and 2*, in which some typical mechanical composition curves are shown.* It will be remembered that in an ordinary mechanical analysis the fractions reported exclude calcium carbonate and organic matter. For convenience of representation the data on

*Equivalent settling-velocities for coarser fractions are calculated.*
which the curves are based have been calculated as percentages of the humus-free mineral soil and thus add up to 100 per cent.

The curves shown in Fig. 1 relate to materials in which, with the exception of kaolin, comminution was known to have been entirely physical.
Curves I and II have been constructed from some data presented by W. Shaw for the percentage of material passing different sized screens in the preparation of road metal.

Curve III relates to a sample of quartz powder obtained through a Weatherhead mortar.

Curve IV relates to a commercial sample of kaolin.

Curve V relates to an air-floated powder, Fullersite, obtained from the grinding of slate waste in the Penrhyn Quarries, near Bangor.

The most remarkable feature of the curves in Fig. 1 is their smooth character, which has been verified also for other mechanically comminuted materials. The general form of the curves is sigmoid, but the upper limb is generally less pronounced than the lower limb. Over the greater part of their range the curves approximate to linear form. The slope of a curve at any point is a measure of the frequency of the fraction represented by the corresponding settling-velocity. At the point of inflection, the curve is steepest and this corresponds with what may be termed the modal fraction. In other words, if the fractions be chosen at equal logarithmic intervals, the largest fraction is shown by the steepest portion of the curve.

It is of interest to note the essential similarity between Curves I and II for road metal with particles up to about 10 cm. in diameter, and Curve V for slate dust in which the coarsest particles have a diameter of about 0.02 mm.

The curves in Fig. 2 refer to natural soils and clays, in which physical and chemical agencies have contributed in varying degrees to their commination. The materials represented are the following:

- Curve I. Sandy red loam, Tanganyika.
- Curve II. Shale loam, Caernarvonshire.
- Curve III. Clay loam subsoil, Hertfordshire.
- Curve IV. Carboniferous subsoil clay, Flintshire.
- Curve V. Clay fraction from aeolian soil, Sudan.
- Curve VI. Red clay, Kenya.
The smooth character of the curves is again evident, but, except in Curve II., there are marked differences from the curves in Fig. 1. The lower limb of the sigmoid appears to be suppressed and the steepest part of the curve is at the lower end of the range. The limit is, of course, obtained by extrapolation, and is therefore conjectural, but appears to lie between log V 7.0 and 7.5. The latter limit would cor-
respond with a particle-diameter of about 0.00006 mm.; but apart from the doubtful validity of Stokes' Law for such small settling-velocities, there are disturbing aqueous sheath effects to which reference will be made at a later point. It is therefore unsafe to take the diameter given as actually representing a lower limit of particle-size.

Considering the curves in detail, they appear to be composite in character. This is most evident in Curves I and III, whose form suggests that these soils are made up of two classes of material, namely, fine material with summation curves wholly convex to the Y-axis, and coarse material with curves of the sigmoid type given by mechanically comminuted material. We may regard the former as being the clay complex formed of the products of chemical weathering and the latter as undecomposed mineral and rock fragments. Curves IV and VI relate to soils consisting almost entirely of clay material, the presence of coarser material being evident only at the upper end of the range. Curve V relates to a clay fraction from which mechanically comminuted material may be presumed to be absent.

Curve II is of interest because it relates to a comparatively juvenile soil in which comminution has been mainly mechanical. It resembles the curves of Fig. 1, but has a greater range, and shows a trend similar to that of the other soil curves at the lower end.

REPRESENTATION OF MECHANICAL COMPOSITION BY TRIANGULAR CO-ORDINATES

Whilst the representation of mechanical composition by means of summation curves has many advantages, notably the information given as to homogeneity of material, for many purposes a system in which mechanical composition is indicated by a single point is preferable. Many workers have proposed the representation of these fractions, sand, silt, and clay, by means of triangular co-ordinates. For example, in the system proposed by J. A. Prescott, J. K. Taylor, and J. J. Marshall, the three fractions are: clay <0.002 mm.; silt 0.002-0.02 mm.; and sand 0.02-2.0 mm.
The disadvantage of this system is that the sand fraction, which includes both coarse and fine sand, is unduly magnified at the expense of the silt and clay.

A more satisfactory separation would be obtained by using the following limits:

<table>
<thead>
<tr>
<th>Particle size, mm.</th>
<th>log. v.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay &lt;0.002</td>
<td>7.5-4.5</td>
</tr>
<tr>
<td>Silt 0.002-0.06</td>
<td>4.5-1.5</td>
</tr>
<tr>
<td>Sand 0.06-2.0</td>
<td>1.5-2.5</td>
</tr>
</tbody>
</table>

Fig. 3.—Mechanical composition represented by triangular co-ordinates.

There is, of course, some uncertainty as to the lower limit of the clay, but in the absence of any more precise indication, log. v = 7.5 is probably not greatly in error, and the suggested separation gives approximately equal intervals.

The fractions on the proposed scale can be readily derived by interpolation from mechanical composition curves.
constructed from results on the International scale. In Table II. are given the mechanical analyses of a number of typical soils belonging to different texture grades, (1) on the International scale and (2) on the suggested three-fraction scale by interpolation. The figures are rounded off to the nearest integer. The results can now be transferred to the triangular diagram, Fig. 3, in which the corners represent 100% sand, 100% silt, and 100% clay, respectively.

In so far as texture can be defined in terms of mechanical composition, we may define sands as soils with more than 50% sand, silts as soils with more than 50% silt, and clays as soils with more than 50% clay. On the diagram these regions are shown by appropriate triangles, whilst the inner triangle indicates the loam textures.

**Table II.—Mechanical analyses of typical soils on International scale and on three-fraction scale.**

<table>
<thead>
<tr>
<th></th>
<th>International Scale</th>
<th>3-fraction Scale</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>C. Sand</td>
<td>F. Sand</td>
</tr>
<tr>
<td>Coarse Sands (Sa)</td>
<td>84</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>52</td>
<td>32</td>
</tr>
<tr>
<td></td>
<td>22</td>
<td>74</td>
</tr>
<tr>
<td>Fine Sands (Sb)</td>
<td>57</td>
<td>24</td>
</tr>
<tr>
<td></td>
<td>41</td>
<td>37</td>
</tr>
<tr>
<td></td>
<td>19</td>
<td>63</td>
</tr>
<tr>
<td>Light Loams (La)</td>
<td>25</td>
<td>33</td>
</tr>
<tr>
<td></td>
<td>20</td>
<td>46</td>
</tr>
<tr>
<td></td>
<td>19</td>
<td>34</td>
</tr>
<tr>
<td>Medium Loams (Lm)</td>
<td>15</td>
<td>35</td>
</tr>
<tr>
<td></td>
<td>16</td>
<td>27</td>
</tr>
<tr>
<td></td>
<td>13</td>
<td>37</td>
</tr>
<tr>
<td>Heavy Loams (Lb)</td>
<td>8</td>
<td>18</td>
</tr>
<tr>
<td></td>
<td>9</td>
<td>28</td>
</tr>
<tr>
<td></td>
<td>9</td>
<td>26</td>
</tr>
<tr>
<td>Silts (Z)</td>
<td>6</td>
<td>28</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>29</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>10</td>
</tr>
<tr>
<td>Clays (C)</td>
<td>2</td>
<td>18</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>0</td>
<td>7</td>
</tr>
</tbody>
</table>
SIGNIFICANCE OF MECHANICAL ANALYSIS AT LOWER END OF PARTICLE-RANGE.

We must now refer to a consideration which affects the significance of the data for the finer fractions in mechanical analysis, namely, the effect of the cation associated with the dispersed particles. Dispersion is usually effected in an alkaline medium such as dilute ammonia or dilute sodium hydroxide solution. According to the cation used, we may refer to an ammonium soil, a sodium soil, or, if the soil has been bereft of associated bases by acid treatment and prolonged washing, a hydrogen soil. Inasmuch as ions vary in their degree of hydration, the particles of a suspension subjected to mechanical analysis will be invested with aqueous shells of varying thickness. It has been shown by G. W. Robinson¹⁰ that, if \( v \) be the settling velocity of a particle of radius \( r \), the effect of an aqueous sheath of thickness \( d \) would be to reduce the velocity to \( vr/(r+d) \). For particles down to the limits of an ordinary mechanical analysis, the aqueous sheath effect is scarcely appreciable, and sodium, ammonium, and hydrogen suspensions yield practically the same results. With particles of smaller size, however, differences in cation hydration lead to different results being obtained with different cations. Comparing a heavily hydrated cation such as sodium with a slightly hydrated cation such as hydrogen, it is evident that a particle associated with sodium ions will carry a larger aqueous sheath than a particle associated with hydrogen ions, and, if the two particles are of equal size, the sodium particle will have a smaller settling-velocity. (See pp. 174 et seq.)

Below the ordinary clay limit, therefore, the mechanical composition for hydrogen soil should give a truer picture than that for sodium soil, in which the aqueous sheath effect tends to make particles settle more slowly and therefore, assuming the ordinary size-velocity relationship, appear smaller than their actual dimensions. C. E. Marshall¹¹ has suggested that clays associated with different cations may actually have different degrees of dispersion and that a particle of sodium
clay is actually smaller than a particle of, say, hydrogen clay. The significance of the lower portion of a mechanical composition curve is uncertain also because of the possibility that the finer fractions thus indicated may not exist as such in the original soil, but originate from the breaking down of relatively larger particles during the pre-treatment.

THE CLAY FRACTION

If the fractions distinguished in mechanical analysis be isolated by using a decantation method it will be seen that they exhibit important differences. Whilst the coarser fractions, namely, the sands and the silt, consist usually and mainly of original fragments such as quartz, felspars, and micas, the clay fraction consists almost entirely of material of secondary origin and of colloidal character. It should be noted, however, that in materials comminuted by mechanical means, although appreciable proportions of the clay fraction occur, it is not in such cases colloidal in character. For example, the slate dust represented by Curve V in Fig. 1 contains 59% of clay fraction, but is lacking in the typical properties of colloidal clay, e.g. plasticity and cohesion.

The colloidal properties which distinguish the clay fraction of soils most markedly from sand and silt may be briefly enumerated. Firstly, colloidal clay exhibits in a high degree the property of absorbing water. This absorption is essentially similar to that shown by such typical colloidal gels as agar and gelatin. The presence of considerable proportions of clay confers a high water-holding capacity on soils and it is to be noted that, whilst the water retained by non-colloidal sand and silt is in the form of surface films, the water associated with clay is held within the cellules of the colloid.

Secondly, changes in the water content of colloidal clay are accompanied by marked changes in volume. This is plainly demonstrated by the cracks which develop in clay soils on drying. Indeed, the shrinkage of highly colloidal
soils may lead to the development of a marked columnar structure whereby the soil breaks up into prisms of polygonal cross-section.

Thirdly, colloidal clay exhibits the property of plasticity when associated with certain proportions of water. In this respect it is sharply distinguished from sand and silt, for even the finest silt is devoid of this property.

Fourthly, colloidal clay confers cohesive properties on soils. If a quantity of silt free from clay be moistened and moulded into the form of a brick, it will be found that the brick crumbles under the slightest stress on drying. Where colloidal clay is present, even in comparatively small proportions, the dried moulded soil exhibits considerable resistance to applied stresses. This is illustrated by the difficulties attending the tillage of clay soils, which, if allowed to dry into clods, are no longer amenable to the ordinary operations of tillage.

Colloidal clay occurs associated with colloidal organic matter and is present in the soil partly as independent aggregates and partly in association with the non-colloidal particles. In this way it renders possible the existence of compound-particle structure and confers on soils the capacity for changes in structure or state of aggregation. Colloidal clay in aqueous suspension is highly sensitive to the presence of small quantities of electrolytes. In the presence of neutral salts it tends to form compound aggregates and is said to flocculate. On the other hand, in weak alkaline suspensions it tends to disperse into its individual particles or, in other words, to deflocculate. Similar changes in aggregation are possible in soils under field conditions. Certain soils containing sodium carbonate are highly impervious to water, and are sticky and intractable under tillage operations, which may even become impossible. In such soils, the individual particles exist independently and there is little or no aggregation into compound particles or crumbs. The question of structure, including crumb formation, is discussed in Chapter VIII.
Finally, the colloidal clay is pre-eminently the reactive constituent of the mineral portion of the soil. It has the typical colloidal property of absorbing certain substances from solution. The absorptive properties of soils, however, reside both in the colloidal clay and in the colloidal organic matter, which are, for most purposes, conveniently considered together as a single complex, namely the *clay-humus complex*. The reactions of this complex will be considered in greater detail in Chapter VI.

Whilst the non-colloidal particles of the soil are almost entirely crystalline fragments of rock-forming minerals, clay was formerly considered to be amorphous. Recent studies, however, by the method of X-ray analysis have shown that most of the clay in soils is actually crystalline and built up of small platelets or fibres. Such a condition is not inconsistent with gel character, for many, if not all, colloidal gels have already been shown to consist of crystalline elements.

In this connexion it is noteworthy that R. Bradfield and others have shown that, for the same soil, the colloidal clay is apparently of uniform composition. A. F. Joseph has shown that the fractionation of colloidal clay by the use of different settling-velocities does not actually reveal distinct fractions, but that the coarsest particles within the clay fraction can be further dispersed. It may be, therefore, that the conception of a range of particles of varying size obtained from a consideration of the non-colloidal fractions is not transferable to the clay fraction and that the coarser grades in this fraction are actually compound particles built up of constituents of similar lattice-structure. This does not exclude the possibility that there may be more than one type of material present.

**ORGANIC MATTER**

Whilst soil exhibits, so far as its mineral portion is concerned, a close resemblance to other unconsolidated products of superficial weathering and even to non-metamorphosed sediments, which are classed geologically as rocks, the pre-
sence of organic matter confers on the soil its distinctive character as the seat of plant life. It is true that some sedimentary rocks contain organic matter; but the organic matter in such cases, though doubtless originally similar to that in the soil, has become considerably modified by secular changes. Whilst the soil organic matter is the habitat of a complex micro-flora and fauna, the organic matter of such sediments as bituminous shales and coal is completely devoid of microbiological activity and, if brought under the influence of soil micro-organisms, is as indifferent to their activities as so much inert mineral matter.

The altered organic matter of such sediments as shales and coals also differs from soil organic matter in the much higher proportion of carbon which it contains. Whilst soil organic matter contains from 55 to 60% of carbon, the organic matter of shales contains much higher proportions; and in the extreme case of anthracites, is almost entirely carbon.

The proportion of organic matter in soil varies considerably. Whilst in certain tropical soils it may fall below 1%, it may constitute practically the whole of the dry matter of peats. The factors influencing the amount of organic matter in soils will be discussed more fully in Chapter VII.

The determination of the exact proportion of organic matter in soil is fraught with considerable difficulty. The loss on ignition is frequently reported as a measure of organic matter; but the figure is almost valueless in the majority of cases since it includes chemically combined water in the mineral portion of the soil and, in calcareous soils, the carbon dioxide driven off from carbonates. And thus it may happen that a heavy soil practically devoid of organic matter may show a loss on ignition of 10 to 12%, consisting almost entirely of chemically combined water driven off at the temperature of ignition. Highly calcareous soils may also show ignition losses out of proportion to the amount of organic matter which they contain. The error in using loss on ignition as a measure of the organic matter is least serious
in the case of sandy soils; but, even here, the determination of small quantities is subject to considerable error.

G. J. Bouyoucos has proposed a method for deriving the proportion of organic matter by subtracting from the ignition loss the combined water, determined by a distillation method.

W. O. Robinson has proposed the estimation of soil organic matter by treating the soil with 15-30% hydrogen peroxide and determining the loss. The objection to this method is that the oxidation is not always completely quantitative, and also that it cannot be assumed that the treatment is without effect on the mineral portion of the soil. The present writer has found that, even with 6% hydrogen peroxide, the oxidation of organic matter is accompanied by a certain solution of mineral matter, principally sesquioxides, consequent on the formation of organic acids. Pyridine and acetyl bromide have been proposed as solvents for organic matter, but they have not as yet come into general use.

The more usual method of estimating the amount of organic matter in a soil is by determining the amount of organic carbon present. This can be carried out either by direct combustion or by one of the methods of wet oxidation. Where the amount of carbon dioxide evolved is measured, it is necessary to apply a correction in the case of calcareous soils, preferably by removing the calcium carbonate in a preliminary treatment. To obtain the amount of organic matter, the figure for organic carbon is multiplied by a conventional factor, usually 1.724, obtained on the assumption that soil organic matter contains 58% of carbon. The use of this factor is open to considerable criticism, for it is based on the composition of the organic matter of peaty soils and not on the composition of the organic matter of ordinary soils. Since no method is known or is likely to be discovered whereby the organic matter of soils can be isolated without alteration, it seems clear that some doubt must always exist as to the exact factor to be used. For comparative pur-
poses, however, the use of the organic carbon itself as an index of organic matter would appear to be the most useful procedure.

The organic matter of the soil consists partly of recognizable fragments of plant material, and partly of material that has lost all traces of its original structure and has become changed into dark amorphous humified matter or humus. Whilst in certain cases, such as the superficial layers of grass soils, the undecomposed material with recognizable structure may form a considerable proportion of the total organic matter, usually 80% or more consists of dark-coloured structureless material that has been derived from added plant residues by a complex series of microbiological and, to some extent, chemical decompositions.

A certain indefiniteness exists in the terminology applied to soil organic matter. Whilst many writers use humus as synonymous with organic matter, others have applied the term to a definite fraction, for example, the so-called matière noire of Grandeau, soluble in 4% ammonia. This confusion has led to the abandonment of the term humus by many writers. In the opinion of the present writer the name humus might be conveniently retained for the organic matter of the soil that has been decomposed and has lost its original structure. Humus would thus exclude recognizable fragments of plant materials that have not yet become part of the characteristic organic matter of the soil. The fact that there is no generally received method for determining humus, as distinct from total organic matter, should not prevent us from recognizing its existence and referring to it by an appropriate name.

Much of what has been said of the colloidal clay applies to humus, which is also of a colloidal character. Its characteristic properties are exhibited in an eminent degree by peat, which is distinguished by its high capacity for absorbing water and also shows great changes in volume on wetting and drying.
Plasticity does not appear to be a distinctive character of colloidal organic matter, whilst cohesion is shown to a much smaller degree than by clay. Indeed, the presence of sufficient humus in the soil may partially neutralize some of the undesirable physical consequences of a high proportion of colloidal clay. At the same time, humic matter confers sufficient cohesion to render possible the formation of compound aggregates or soil crumbs in sandy soils and to that extent exerts a favourable effect.

Humus is requisite for the micro-biological processes of the soil and is, with its associated moisture, the medium in which soil organisms fulfil their life-cycles. The presence of humus and its living population is the distinctive character of a soil, for, whilst raw subsoils contain colloidal clay, and even small amounts of organic matter, they are, below moderate depths, devoid of living organisms.

The great complexity of soil organic matter has long been realized. In Chapter VII we shall discuss its constitution in connexion with the decomposition of plant residues in the soil. We may note, however, that the greater part of the soil humus consists of relatively inert material whose rôle in the soil is mainly physical. The material directly concerned in micro-biological changes includes undecomposed plant residues, and a comparatively small fraction of the humus, consisting of simple and presumably active compounds.

THE CLAY-HUMUS COMPLEX

For convenience of presentation, we have, up to the present, considered the clay and the humus separately. From the point of view of the constitution of the soil, however, they cannot be considered apart, since they are associated together, if not in actual chemical combination, in the most intimate physical relationship, and form the clay-humus complex of the soil. Whilst the non-clay mineral matter of the soil exhibits in some degree those properties of colloids associated with surface development,
it is in the clay-humus complex that these properties are principally seated.

Since clay and humus differ considerably in their properties, it is obvious that the gross amount of colloidal matter present is no true measure of the colloidal nature of the soil. A soil containing 25% of colloidal material, made up of 20% of clay and 5% of organic matter, will differ markedly from a soil also containing 25% of colloidal matter, made up of 5% of clay and 20% of organic matter. Data purporting to give the percentage of "colloid" in the soil should be interpreted with caution.

The colloidal constituent of soils occurs usually in films or coatings investing the non-colloidal particles. In highly colloidal soils, however, it is probably present also in the form of independent aggregates in which, doubtless, fragments of macro-crystalline material may be embedded. It is unusual to find soils entirely composed of colloidal matter, but certain highly weathered tropical soils may contain as much as 90%, of which nearly the whole is colloidal clay. The non-colloidal material in such soils consists of unweatherable minerals such as quartz and magnetite. In extreme cases, peat soils may consist almost entirely of organic matter, of which the greater proportion is colloidal.

It is worthy of note that when soils are prepared for laboratory examination by light grinding and sieving, the larger aggregates in the sample contain more colloidal clay and organic matter than the finer material. For example, A. M. Smith\textsuperscript{16} found that aggregates larger than 1 mm. in diameter in a 3 mm. sample of soil had a greater content of exchangeable bases than had the finer material.

SOIL SOLUTION

The study of the soil solution offers special difficulties owing to the fact that the soil cannot be regarded simply as a mixture of insoluble inert material with substances of vary-

\*A sharp boundary between non-colloidal core and colloidal film cannot be assumed, but rather, a gradual transition associated with increasing hydration.
ing solubility. For example, in the case of a mixture of quartz sand and calcium sulphate, the concentration of the liquid phase in equilibrium would remain constant as long as any calcium sulphate remained undissolved. Thereafter, the concentration would vary inversely as the volume of the liquid phase. The situation in the soil is different because of the reactivity of the insoluble solid phase, which exhibits the phenomena of absorption and base exchange. (See Chapter VI).

At the dilutions ordinarily encountered in the soil solution, the salts present are almost entirely dissociated into their component ions. In considering the equilibria between the solid phase of the soil and the soil solution, it is therefore necessary to think in terms of ions rather than salts. Thus, the soil solution has a certain concentration of calcium-ion and this concentration will vary with the ratio of soil to water. The soil calcium with which the calcium of the soil solution is in equilibrium is the exchangeable calcium, combined with the insoluble acids of the clay and organic matter.

Certain ions, notably chloride, and nitrate, are not ordinarily absorbed by soils and there is, thus, no soil chloride, or soil nitrate with which they are in equilibrium. The concentration of these ions in the soil solution, in the absence of any removal by leaching or root absorption, will therefore be inversely affected by the ratio of water to soil.

The above considerations should make it clear why the soil solution cannot be investigated simply by adding a known volume of water to a given weight of soil, filtering, and analysing an aliquot of the filtrate for the purpose of calculating back to the original water content of the soil. It has therefore been necessary to seek for methods whereby the actual solution can be isolated from soil at moisture contents comparable with those occurring under field conditions.

In one group of methods the moist soil is centrifuged at high speed, whereby a certain proportion of the soil solution
is thrown out. By no means the whole of the soil solution is thus removable and the method is of only limited application. Ordinary pressure methods are of equally limited use.

The most convenient methods for obtaining the soil solution depend on displacement by other liquids. Moist soil is carefully packed into cylinders with perforated bottoms and the surface of the soil is flooded with a suitable liquid. This percolates into the soil, driving the soil solution before it. Eventually the lower part of the soil column becomes sufficiently saturated with the displaced soil solution for an excess to drain away. Drainage and collection are allowed to continue until the displacing liquid itself begins to percolate through. In the original experiments of Schloesing in 1866, water coloured with carmine was used, the appearance of the dye in the percolate indicating the point at which the original soil solution had been displaced. Paraffin oil and ethyl alcohol have also been used.

The results expressed as parts per million, obtained by Schloesing for a soil containing about 20% of moisture, are as follows:—

<table>
<thead>
<tr>
<th>SiO₂</th>
<th>N₂O₅</th>
<th>CO₂</th>
<th>CaO</th>
<th>MgO</th>
<th>K₂O</th>
<th>Na₂O</th>
<th>SO₃</th>
<th>Cl</th>
<th>Organic Matter</th>
</tr>
</thead>
<tbody>
<tr>
<td>29.1</td>
<td>395</td>
<td>118</td>
<td>264</td>
<td>13.5</td>
<td>6.9</td>
<td>7.8</td>
<td>57.9</td>
<td>7.4</td>
<td>37.5</td>
</tr>
</tbody>
</table>

J. S. Burd and J. C. Martin have examined the soil solution of certain Californian soils, using a method of displacement under pressure. Some results for a soil at the beginning and end of the crop season are given in Table III.

**Table III.—Composition of soil solution (Burd and Martin).**

<table>
<thead>
<tr>
<th>Date</th>
<th>pH</th>
<th>Parts per million in displaced solution</th>
<th>Total Solids</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>NO₃</td>
<td>HCO₃</td>
</tr>
<tr>
<td>April 30th, 1923</td>
<td>7.4</td>
<td>149</td>
<td>83</td>
</tr>
<tr>
<td>Sept. 4th, 1923</td>
<td>7.6</td>
<td>58</td>
<td>155</td>
</tr>
</tbody>
</table>
The results are of the same order of magnitude as those obtained by Schloesing, the most notable difference being the higher sodium content in the solutions from the Californian soil. The decrease in concentration consequent on the growth of a crop is to be noted.

The figures given merely serve to indicate the order of magnitude of the concentration of the principal ions. It may be expected that the concentration of the soil solution will increase with decreasing moisture content, and will be greater in dry climates than in humid climates, where soluble materials are leached out by percolating waters and lost in drainage.

There are undoubted difficulties in the way of displacement methods, which in practice prove to be of limited application. The moist soil must be well packed in the displacement cylinder in order to avoid percolation of the displacing liquid through wide capillaries. This is an obstacle in the way of investigation of heavy soils since any attempt at packing results in complete hindrance of percolation.

Recently, R. F. Reitemeier and L. A. Richards have developed an ultra-filtration method in which soil in contact with a hydrophilic cellulose membrane supported on 100-mesh brass wire cloth is subjected to a gas pressure of up to 15 atmospheres. The other side of the membrane being at atmospheric pressure, moisture is forced out from the soil and through the membrane. The results were in good agreement with those obtained by displacement methods for a number of soils, and it may be surmised that for those soils where displacement methods are inapplicable, e.g. heavy clays, ultra-filtration might prove a satisfactory method. It is improbable that the composition of the soil solution is uniform throughout a given soil and appreciable ionic concentration gradients may exist within soil moisture films. It is therefore unlikely that a complete picture of the soil solution has yet been obtained.

Drainage waters, though more dilute than the soil
solution, show a general resemblance in composition. Abundant data are available, but perhaps the most instructive are those obtained from the Broadbalk wheat field at Rothamsted, where drainage is collected from plots which have been continuously under wheat for about 80 years. Table IV gives the data obtained by A. Voelcker for drainage from three of the plots. These soils contain carbonates. With acid soils, the lime figures would be lower.

**Table IV.—Analyses of Drainage Water from Cultivated Fields, Parts per Million (A. Voelcker)**

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Unmanured</th>
<th>Farmyard manure</th>
<th>Complete artificial fertilizers</th>
</tr>
</thead>
<tbody>
<tr>
<td>CaO</td>
<td>98.1</td>
<td>147.4</td>
<td>143.9</td>
</tr>
<tr>
<td>MgO</td>
<td>5.1</td>
<td>4.9</td>
<td>7.9</td>
</tr>
<tr>
<td>K₂O</td>
<td>1.7</td>
<td>5.4</td>
<td>4.4</td>
</tr>
<tr>
<td>Na₂O</td>
<td>6.0</td>
<td>13.7</td>
<td>10.7</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>5.7</td>
<td>2.6</td>
<td>2.7</td>
</tr>
<tr>
<td>Cl</td>
<td>10.7</td>
<td>20.7</td>
<td>20.7</td>
</tr>
<tr>
<td>SO₂</td>
<td>24.7</td>
<td>106.1</td>
<td>73.3</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.6</td>
<td></td>
<td>1.54</td>
</tr>
<tr>
<td>SiO₂</td>
<td>10.9</td>
<td>35.7</td>
<td>24.7</td>
</tr>
<tr>
<td>N as NH₃</td>
<td>0.14</td>
<td>0.20</td>
<td>0.24</td>
</tr>
<tr>
<td>N as Nitrate</td>
<td>15.0</td>
<td>62.0</td>
<td>32.9</td>
</tr>
<tr>
<td>Org. Matter, CO₂ etc.</td>
<td>67.7</td>
<td>77.3</td>
<td>84.6</td>
</tr>
<tr>
<td>Total Solids</td>
<td>246.4</td>
<td>476.0</td>
<td>407.6</td>
</tr>
</tbody>
</table>

**SOIL AIR.**

Like the soil solution, the soil air is a dependent phase and its composition varies even in the same soil. The chief respects in which it differs from the free atmosphere are: (1) it contains a much larger proportion of carbon dioxide; (2) it is, except in air-dry soils, saturated with water vapour; and (3) it contains rather less oxygen and nitrogen,
The highest proportions of carbon dioxide are found in soils rich in organic matter. For example, the air of grassland soils may contain as much as 1.5% of this constituent. Arable soils may contain up to 0.5%, the highest figures being found in soil recently treated with organic manures. In addition to the free air of soils, there is also air dissolved in the soil moisture and associated with the soil colloids. This consists mainly of carbon dioxide and nitrogen.

In the same soil, the proportion of carbon dioxide in the soil air will vary with the energy of micro-biological decomposition. In Britain, where the soil generally contains sufficient moisture for micro-organisms to fulfil their functions, micro-biological activity in a given soil is controlled mainly by temperature and we should therefore expect the greatest concentration of carbon dioxide to be found during the summer months. Aeration also plays a part in the rate of production of carbon dioxide, since the oxidative decomposition of organic matter is an aerobic process. The amount of carbon dioxide in the soil will not, of course, depend simply on the rate at which it is produced by microbiological activity, for diffusion will also enter as a factor.*

It has been suggested that the rate of evolution of carbon dioxide from the soil is a measure of biological activity and hence of fertility. Whilst this may be true in comparing soils of similar constitution under similar conditions, the biological production of carbon dioxide is governed by so many factors that it is scarcely possible that it can be used as a general measure of fertility.

The carbon dioxide produced in the soil from the biological decomposition of organic matter and from the respiration of plant roots constitutes the principal agency whereby the supply of this constituent, depleted by photosynthesis, is regenerated in the atmosphere. It is also of importance as a constituent of soil moisture and of percolating waters, and thus contributes to the formation of soil

by intensifying the hydrolytic action of water on mineral silicates.

PORE-SPACE

We have seen that the principal constituents of the soil are (1) mineral matter, consisting of particles varying in size from stones to clay; (2) organic matter, consisting of decayed and decaying residues of vegetation and associated with clay to form the colloidal complex; (3) soil solution, which is a dilute solution of the bicarbonates, sulphates, chlorides, nitrates, phosphates and silicates of calcium, magnesium, potassium, sodium and iron; and (4) soil air, which differs from atmospheric air in being saturated with water vapour and enriched in carbon dioxide. It now remains to notice the structure of the soil.

Soil being a granular substance, a given volume of soil will include a certain proportion of interstitial spaces or pore-space, which may be filled either with water or air. We may therefore distinguish the apparent volume of a certain quantity of soil from the true volume occupied by the particles which compose it.

If 100 grams of soil occupy a volume, $V$ c.c., then the apparent density, $d_1$, is $100/V$. Now, if the true density of the soil be $d$, its true volume is $100/d$ and the volume occupied by the interstitial space is $V - 100/d$ or $100/d_1 - 100/d$, which, expressed as a percentage, is $100(d-d_1)/d$.

If a soil could be imagined to consist of uniform spherical particles with closest packing, the pore-space would be 26%. In practice, this condition is never fulfilled, for not only are there particles of different shapes and sizes present, but, owing to the presence of colloidal material, the ultimate particles are to a large extent aggregated together to form compound particles or soil crumbs, so that the pore-space is actually never so low as in the case of uniform spheres with close packing. It generally varies from 40 to 60%.

The pore-space is not a specific property of a given soil, but depends on its temporary structure. Cultivation has the
effect of increasing pore-space, whilst, on the other hand, it may be diminished by pressure or treading. In ordinary cultivated soils the pore-space is at maximum after ploughing and gradually diminishes as the soil settles down under the influence of weather. The effect of frost is to increase pore-space.

The proportion of pore-space in a soil depends on the extent to which its primary particles are united together into compound particles or crumbs. This possibility is greatest in soils containing large proportions of clay or organic matter and least in barren sands and gravels. Inasmuch as a high pore-space is favourable to aeration and water movements, it is obvious that the crumb structure is most desirable for plant growth. On the other hand, the destruction of crumb structure and the prevalence of the so-called single grain structure lead to a diminution of pore-space and unfavourable conditions for plant growth.

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CHAPTER III

SOIL FORMATION: I. THE PARENT MATERIAL

In considering the pedogenic processes, it must be borne in mind that two groups of processes are involved, namely (1) weathering and (2) profile development. The soil profile is developed from parent material that has itself originated from the weathering of rocks. In the simplest case, weathering precedes profile development, as in soil formation from alluvia and unmetamorphosed sediments. The initiation of profile development in weathered material does not, however, put a stop to weathering, and the two processes can proceed simultaneously. The weathering processes cease only when the soil material no longer contains unweathered but weatherable minerals. On the other hand, in many shallow rocky soils such as those of our uplands, soil formation begins and proceeds pari-passu with weathering.

In the present chapter we shall consider those processes whereby the solid rocks of the earth's crust are broken down to form the parent material of soils. This implies the study of weathering, but some account must also be given of the processes whereby weathered products are redistributed by agencies of transport. The subject is as wide as dynamic geology itself, but must be included, although in abbreviated treatment, in any account of the processes of soil formation. In the main we are concerned with the processes of rock weathering consequent on the action of atmospheric agencies on the superficial layers of the earth's crust. Processes occurring at depth, such as deep-seated hydrothermal effects, are only of indirect consequence for our subject.

The rocks of which the earth's crust are composed, when subjected to the action of the atmosphere, undergo decay.
The products of this decay form the regolith, the unconsolidated débris mantling the solid rocks that lie beneath. In the absence of agencies of transport, this mantle would grow in thickness until the underlying rocks would be protected from further attack. By the action of gravity, running water, moving ice, or wind, the products of decay are continually being moved from their place of origin and redistributed on the earth's surface. In this way, fresh layers of the earth's crust are brought within the reach of atmospheric effects. In the absence of any changes in the relative levels of land and sea, the products of weathering are carried down in rivers and accumulate as sediments, which in time become consolidated to form sedimentary rocks. This denudation is terminated when the existing land sinks below sea-level and former sediments, now consolidated into rocks, emerge as land to undergo another cycle of weathering and denudation. Both the regolith and the terrigenous sediments derived therefrom are traced back ultimately to the weathering of crystalline rocks.

**PHYSICAL WEATHERING**

Two main types of weathering may be distinguished, namely (1) physical or mechanical weathering and (2) chemical weathering, the former involving only comminution without chemical change, and the latter involving chemical change with formation of new products. The processes whereby rocks are physically weathered are familiar to students of geology. They are essentially processes which result merely in a comminution of the materials on which they operate. Granite that has only been subjected to mechanical weathering is still granite, except in so far as the space-mosaic of constituent minerals has been broken up. There is no question of chemical change or of the formation of new products.

The most universal type of physical weathering is that produced by changes in temperature. Where such changes are of such sufficient magnitude and suddenness, expansion or contraction in superficial layers of rocks can result in
SOIL FORMATION: I. THE PARENT MATERIAL

strains, which ultimately lead to shattering. This type of weathering is most marked in dry climates, in which free radiation results in great changes in temperature at sunrise and sunset. The weathering of rocks in deserts is predominantly of this type.

In cool or temperate climates with abundant precipitation, the alternations of frost and thaw are of chief importance in the physical comminution of rocks. This action depends on the occurrence in rocks of joints and interstices which, during winter, become filled with water. The serees of our mountain valleys are striking evidences of the disruptive action of alternating frost and thaw, and the same processes are continued in every fragment which, by its porosity, is capable of holding water.

Other agencies of physical weathering are moving water in rivers, moving ice in glaciers, wave action, and the blast of sand grains transported by wind. For a detailed study of these processes the student may be referred to any textbook of geology, to which their study more particularly belongs.

Soils in the early stages of development, desert soils, and the soils of the arctic and alpine regions are predominantly the result of physical weathering. They are generally characterized by the presence of large proportions of coarse material, and their mechanical composition curves often show an approximation to a linear form in their upper range. Soils derived from physically weathered material are often termed skeletal soils.

CHEMICAL WEATHERING

The chemical weathering of rocks involves more serious alterations than the mere comminution produced by physical weathering. As a result of these changes, certain minerals disappear wholly or partially, and material of secondary origin is formed, which differs markedly from the parent minerals.

The processes embraced under this head fulfil themselves in an aqueous medium and depend on the decompos-
ing action of water, reinforced by the presence of dissolved carbon dioxide and, in some cases, organic acids formed from the decay of plant residues. Sulphuric acid formed by the oxidation of organic sulphur compounds or iron sulphides may also play a part. Since chemical weathering takes place at the surface of rock minerals, it is evident that it is intensified where physical comminution has preceded it. But since most rocks are formed of an irregular space-mosaic of different minerals with different degrees of susceptibility to attack, chemical weathering alone is capable of producing disintegration, particularly where there is a certain degree of jointing or porosity in the weathering rock.

Essentially, chemical weathering involves two phases, namely, (a) the disappearance of certain minerals, and (b) the formation of secondary products. Some of the secondary products may originate by alteration in situ of the parent minerals, whilst other products may originate by precipitation from solutions containing the soluble products of weathering. Such precipitation may occur at the seat of weathering, or apart from the seat of weathering after transport by moving water. Material precipitated at the seat of weathering may be mixed, or even enter into combination, with residual products.

The processes may be represented diagrammatically as follows:—

```
ROCK MINERALS

  Residual or alteration products
  /\                             /\      \
 /\                             /\      \
Solution                  Solution

                  /\                           /\                \
Constituents precip-    Constituents precip-            \
itated from solution    itated from solution

                  /\                           /\                \
At seat of weathering   Apart from seat of weathering

Weathering-complex   Illuvial material
```

(see p. 78)
SOIL FORMATION: I. THE PARENT MATERIAL

The weathering-complex, representing that portion of the soil which is of secondary origin, may include both residual or alteration products and material precipitated in situ from soluble products of chemical weathering. It may also include material transported from elsewhere and reprecipitated.

The chemical weathering of rocks affects principally the mineral silicates, such as felspars, micas, and ferromagnesian minerals, and depends on their instability at ordinary temperatures in the presence of water and carbon dioxide. From data by F. W. Clarke, it is estimated that the approximate proportions of the principal minerals in the earth's crust are as follows:—Felspars 57.8%; Amphiboles and Pyroxenes (ferromagnesian minerals) 16.0%; Quartz 12.7%; Micas 3.6%. It is thus evident that the felspars are by far the most important group of soil-forming minerals, followed by the ferromagnesian minerals. Certain minerals are practically unaffected by chemical weathering and persist unaltered in the soil. Of these the principal example is quartz. Other minerals resistant to chemical weathering are magnetite, titanite, ilmenite, etc.

The chemical weathering of rocks depends on the presence of water. Since the rate of chemical change is affected by temperature, it will be recessive at low temperatures and greatest at high temperatures. Translated into terms of natural conditions, this means that chemical weathering is at minimum under desert conditions, owing to absence of water, and under arctic and alpine conditions, owing to the prevalence of low temperatures. In such cases the weathering which occurs is almost entirely of a physical character. In humid climates, the intensity of weathering will increase with temperature and reach a maximum in the tropics.

Soils predominantly of the physically weathered type may occur in a region favourable for chemical weathering. Such soils are generally young soils derived from crystalline rocks and are developed on recent deposits or in situations where erosion is sufficiently active to prevent the accumula-
tion of the products of decay and the consequent predominance of chemical over physical weathering.

SIMPLE SOLUTION

Simple solution, involving the total removal of the weathering material, is negligible except in the case of calcium carbonate which, compared with silicate minerals, is very soluble in percolating waters, particularly when such waters contain dissolved carbon dioxide that has originated from the respiration of plant roots and the decomposition of organic residues in the soil.

The most spectacular effects of the solution of calcium carbonate are seen in the cave systems that are so characteristic of limestone countries. From the point of view of soil formation, the solution of calcium carbonate is of interest by reason of the fact that limestones generally contain an admixture of non-calcareous material, which remains behind as a residue. Limestone soils are thus essentially residual soils, and their character depends principally on the nature of the material associated with the calcium carbonate in the parent limestone. Thus we may have soils of all textures from sands to clays originating from the weathering of calcareous rocks.

Soil formation from limestones cannot always be explained as a simple removal of calcium carbonate. During the process of liberation of the residual material certain changes may occur that profoundly modify its character. The terra rossa soils associated with limestones in the Mediterranean region represent material that has been considerably altered from the original residue. (See Chapter XV.)

HYDRATION AND OXIDATION

Although the silicate minerals are considerably less soluble than calcium carbonate, yet solution, or, at least, the presence of water, is a necessary circumstance in their chemical decomposition. Before dealing with the most im-
important type of chemical weathering of silicates consequent on solution, we must notice two processes in which, although water is necessary for their fulfilment, actual solution is not demonstrably required. *Hydration* consists in the chemical combination of water with a particular mineral. An example of this is the formation of serpentine from olivine. In itself, hydration is not essentially a pedogenic process since it occurs also under conditions remote from the operation of ordinary atmospheric agencies. It may, however, be of importance as a preliminary to the more characteristic pedogenic processes.

**Oxidative processes**, broadly speaking, affect only ferrous compounds. Certain ferromagnesian compounds contain ferrous iron, which, in the presence of air and moisture, is capable of being oxidized with the production of hydrated ferric oxide, e.g.,

\[2\text{FeSiO}_3 + 3\text{H}_2\text{O} + \text{O} = 2\text{Fe(OH)}_3 + 2\text{SiO}_2\]

Iron pyrites and marcasite yield by oxidation hydrated ferric oxide and sulphuric acid:

\[2\text{FeS}_2 + 7\text{H}_2\text{O} + 15\text{O} = 2\text{Fe(OH)}_3 + 4\text{H}_2\text{SO}_4\]

**HYDROLYSIS**

We may now consider the most important of the processes whereby minerals are weathered chemically, namely, *hydrolysis*. Hydrolysis is a consequence of the partial dissociation of water into hydrogen-ions and hydroxyl-ions. The actual amount of dissociation is exceedingly small, for at ordinary temperatures a litre of water contains only $10^{-7}$ g. of hydrogen in the ionic form; but the presence of carbon dioxide in solution increases the hydrogen-ion concentration, and thus reinforces the hydrolytic action of the water. The hydrogen-ion concentration of the soil solution and, consequently, its hydrolytic activity may be further augmented by the presence of organic and mineral acids.

The presence of hydrogen-ions is characteristic of acid solutions. Strong acids such as hydrochloric and sulphuric
acids in dilute solutions are almost completely dissociated. Thus, in a decinormal solution of hydrochloric acid, practically all the hydrogen is in the ionic form, and the hydrogen-ion concentration is nearly 0.1 g. per litre. In a weak acid, on the other hand, dissociation is only partial. A decinormal solution of acetic acid is only slightly dissociated and the hydrogen-ion concentration is about 0.0013 g. per litre.

Water thus behaves as a weak acid, and its effect on weatherable silicate minerals depends on the activity of the hydrogen-ions. The minerals with which we are concerned may be regarded as complex silicates of calcium, magnesium, potassium, sodium, aluminium, and iron. The simplest type of decomposition would consist in the replacement of alkali or alkaline earth-ions in the mineral lattice by hydrogen-ions, resulting in the formation of aluminosilicic or ferrosilicic acids and the liberation of alkali or alkaline earth hydroxides. A further stage in breakdown might consist in the separation of silicic acid, accompanied by a profound modification of the original crystal lattice. A process such as the following may be conceived to occur in the formation of kaolinite from potash felspar:

(1) \[ K_2O \cdot Al_2O_3 \cdot 6SiO_2 + 2H_2O \rightarrow H_2O \cdot Al_2O_3 \cdot 6SiO_2 + 2KOH \]

followed by:

(2) \[ H_2O \cdot Al_2O_3 \cdot 6SiO_2 + H_2O \rightarrow Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O + 4SiO_2 \]

The action of water on felspar is illustrated by the classical experiment of Daubrée, in which 3 kg. of orthoclase felspar, after 200 hours' agitation with water, yielded 2.52 g. of \( K_2O \) in solution and an insoluble alteration product probably of a kaolin-like character. Under laboratory conditions, however, the hydrolysis of silicates, as evidenced by the development of an alkaline reaction, slows down after its initial stages owing to the accumulation of a protecting layer of the insoluble products of the reaction.

O. Tamm, by grinding a potash felspar in water for long periods, obtained a product resembling muscovite mica. It is possible, therefore, that mica represents a stage in the
breakdown of felspars. P. A. Zemiatnchensky obtained similar results.

Much of the earlier literature on weathering was devoted to theories to account for the formation of kaolin from felspar. It is now known that kaolin, or kaolinite, far from being the characteristic ingredient of the weathering-complex of soil may be present in some cases in insignificant proportions. On the other hand, the examination of soils and clays by X-ray methods has revealed the presence of a number of characteristic minerals of known crystal constitution. It is also known that the characteristic minerals of the weathering-complex may contain iron as well as aluminium. Future work will doubtless be directed towards elucidating the mode of origin of these minerals from the primary minerals. It will be necessary, in the first place, to decide whether a given clay mineral is formed as a residual product, or whether it is formed by precipitation from solution. It may indeed be possible to classify the minerals of the weathering-complex into residual minerals and precipitation minerals.

A considerable body of experiment and theory has centred round the origin of the weathering-complex by precipitation from solutions. According to the precipitation hypothesis, the products of chemical weathering are formed by precipitation from the soluble products of silicate hydrolysis.

Although the original silicate minerals are highly insoluble, the consequences of such slight solubility as they possess may be appreciable over long periods of time.

It is supposed that the first stage consists in the simultaneous liberation of silicic acid and the hydroxides of the bases present. Whilst the alkali and alkaline earth hydroxides are soluble in water, the silicic acid and the hydroxides of aluminium and iron may form colloidal solutions. Silicic acid sols are electronegative, whilst the sesquioxide sols are electropositive, and mutual precipitation may occur, giving rise to the weathering-complex. This may take place in one of the following ways:
(a). The mutual precipitation may, as originally suggested by J. M. Van Bemmelen, result in the formation of amorphous absorption compounds of indefinite composition. The general formula for such compounds might be written,

\[ \text{a SiO}_2 \cdot \text{b Al}_2\text{O}_3 \cdot \text{c Fe}_2\text{O}_3 \cdot \text{d CaO} \cdot \text{e MgO} \cdot \text{f K}_2\text{O} \cdot \text{g Na}_2\text{O} \cdot \text{h H}_2\text{O} \]

(b). The absorption compounds of indefinite composition and amorphous structure initially formed may give rise to crystalline compounds of definite, though complex, composition.

(c). The mutual precipitation may give rise directly to crystalline compounds of definite, though complex, composition.

S. Mattson studied the precipitates formed under different conditions from solutions analogous to those that might be expected to produce the weathering complex. In these experiments standard solutions were mixed in varying proportions and the flocculation was observed immediately, and after standing overnight. In addition, the cataphoretic transport (see p. 179) under a definite potential gradient was determined ultramicroscopically. Precipitates showing no cataphoretic transport are termed isoelectric and correspond with maximum precipitation.

Data have been obtained for the formation of precipitates in which the basic and acidic components are as follows:

<table>
<thead>
<tr>
<th>Basic</th>
<th>Acidic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Al₂O₃</td>
<td>SiO₂</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>P₂O₅</td>
</tr>
<tr>
<td>Humic acid</td>
<td></td>
</tr>
</tbody>
</table>

By carrying out experiments in the presence of hydrochloric acid or sodium hydroxide, the effect of reaction on isoelectric precipitation was also studied.

The following results were obtained in a typical experiment.
SOIL FORMATION: I. THE PARENT MATERIAL

TABLE V.—Flocculation of Aluminium Chloride and Sodium Silicate Solutions (Mattson).

<table>
<thead>
<tr>
<th>C.C. Solution A</th>
<th>C.C. Solution B</th>
<th>Flocculation</th>
<th>Transport Velocity</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>20</td>
<td>17.4</td>
<td>opalescent</td>
<td>O</td>
<td></td>
</tr>
<tr>
<td>20</td>
<td>17.6</td>
<td>slow</td>
<td>XXXX²</td>
<td>+1.38</td>
</tr>
<tr>
<td>20</td>
<td>17.8</td>
<td>instant</td>
<td>XXXX</td>
<td>+0.76</td>
</tr>
<tr>
<td>20</td>
<td>18.0</td>
<td>instant</td>
<td>XXXX</td>
<td>−0.67</td>
</tr>
<tr>
<td>20</td>
<td>18.5</td>
<td>slow</td>
<td>XX</td>
<td>...</td>
</tr>
</tbody>
</table>

A = 5 millimoles AlCl₃ per litre. B = 7.5 millimoles Na₂SiO₃ per litre.
Molecular ratio in isoelectric mixture: SiO₂/Al₂O₃ = 2.68
pH at isoelectric point: = 6.25.
Composition of isoelectric precipitate: SiO₂/Al₂O₃ = 1.63.

*Complete flocculation indicated by XXXX.

The precipitate with 20 c.c. of Solution A and 17.8 c.c. of Solution B is electropositive, as is shown by its direction of transport, whilst the precipitate with 20 c.c. of Solution A and 18.0 c.c. of Solution B is electronegative. The isoelectric precipitate, ex hypothesi, shows no cataphoretic movement. Precipitation is most complete at the isoelectric point; but it should be noted, from the discrepancy between the molecular ratio in the isoelectric mixture and in the isoelectric precipitate, that precipitation is incomplete. In the above experiment an excess of silicic acid would remain in solution.

Mattson’s results for the different systems obtainable from alumina and ferric oxide on the one hand, and silica, phosphoric oxide, and humus on the other, are shown diagrammatically in Fig. 4.

There is, for every pH, a corresponding isoelectric precipitate. It will be seen that for the SiO₂-Al₂O₃ system, the molecular ratio in the isoelectric precipitates varies from about 2.4 at pH 4.5 to zero at pH 8.0. In the case of the SiO₂-Fe₂O₃ system, the range of molecular ratios is from about 2.1 at pH 5.0 to zero at pH 7.0. In the presence of
bivalent cations—for example, calcium—more siliceous complexes are obtained.

The colloidal complex of the soil, according to Mattson, may be considered to result from mutual precipitation, at or near isoelectric conditions, from electropositive (basic) sols on the one hand and electronegative (acidic) sols on the other.

**Figure 4.**—Relation between pH and isoelectric composition of aluminium and ferric "phosphates," "silicates," and "humates." (S. Matison.)

(From "Soil Science"—by permission.)
Whilst it is scarcely possible that sols of opposite sign could originate simultaneously as the result of weathering, and then form isoelectric precipitates as in laboratory experiments, the isoelectric precipitate theory may serve as a useful hypothesis to explain a process in which the intermediate steps postulated by this theory are suppressed, for the isoelectric precipitate represents the most stable product under the given conditions.

In later papers S. Mattson and Y. Gustafsson have developed their theory of isoelectric weathering. They regard the material of the soil colloidal complex, which we may equate with the weathering complex, as existing in seven distinct forms according to the following scheme:—

All degrees of transition between these forms occur. Positive and negative charges are balanced at the isoelectric point. As the pH of the medium is lowered or raised the gel complex becomes increasingly cationic or anionic re-
spectively and at a certain point goes over into the sol condition. The complex becomes not only more highly dispersed but undergoes hydrolytic decomposition. The molar ratios of silica to sesquioxides in parent rocks are considerably higher than those found for the weathering complex. If the first product of weathering should correspond in composition with the parent rock, its isoelectric point would be very low, particularly if humus is present in addition to silicic acid as an acidic constituent. Since parent rocks contain relatively high proportions of strong bases, which become liberated during weathering, the pH of the weathering material may be expected to be above the isoelectric point of the complex. Under these conditions an anionic sol complex is liberated whilst the gel complex will tend to become richer in basoids with consequent increase in its isoelectric pH. This would continue until the isoelectric point of gel complex approximates to the pH of the weathering medium.

It is difficult to regard the mutual precipitation theory otherwise than as a helpful fiction. If mutual precipitation is to occur, then the positive sesquioxide sols and the negative silicic acid sols must first achieve separate existence. Can we imagine that, at one point in the weathering material, sesquioxide sols are formed, whilst at another point the corresponding silicic acid sols originate? The sesquioxides and silicic acid sols must at the moment of liberation be as closely adjacent as their positions in the parent lattice demand, and they can have no possibility of forming independent sols. Nevertheless, considering the isoelectric precipitate to be the most stable system in equilibrium with the liquid phase under the conditions of experiment, the assumption of a separate existence for the positive and negative sols may be dispensed with. Mutual precipitation, as such, is not an essential feature of Mattson’s theories. The general agreement of his deductions with the observed facts under different conditions of weathering lends strong support to their general validity as a working hypothesis. Mattson’s theories may be equally
applicable to a theory of hydrolysis which conceives the first stage to consist in the formation of residual aluminosilicic and ferrosilicic acids.

GENERAL RESULTS OF HYDROLYSIS

However we may conceive the process of hydrolysis, certain general consequences may be noted:—

(1) De-silicification occurs. Comparison of weathered materials with their parent rocks generally reveals a loss of silicic acid in the weathering process. The silicic acid is removed by percolating waters, to a great extent in the form of silicates of the alkali and alkaline earths. It is significant that river waters, which contain the material lost by solution in the weathering processes, contain an excess of silicic acid over sesquioxides. In arid and semi-arid regions incomplete leaching may prevent the removal of silicic acid liberated by hydrolysis. In such cases the weathering-complex is characterized by a high content of silica.

(2) De-alkalization occurs during hydrolysis. Comparison of fresh rocks with their weathered products reveals a general loss of calcium, magnesium, potassium, and sodium. Exceptions to this rule occur in arid regions, where alkali and alkaline earth salts may accumulate, giving rise to saline or alkaline soils.

(3) Hydrolysis results in the formation of new substances, either by modification of the original minerals or by a partial re-synthesis of the products of decomposition. These substances are collectively referred to as the weathering-complex, the zeolitic complex, the clay complex, or the inorganic soil colloid. They constitute the greater part of the clay fraction in mechanical analysis and are almost entirely found in that fraction. The study of the clay fraction, therefore, offers the most convenient basis for the elucidation of the composition and constitution of the weathering complex.

The following data from G. E. Merrill illustrate the losses involved in chemical weathering (Table VI).
Table VI.—Analyses of Fresh and Decomposed Diorite. Albemarle County, Virginia (Merrill).

(From “Rocks, Rock Weathering, and Soils,” by permission of The Macmillan Co., Publishers.)

<table>
<thead>
<tr>
<th></th>
<th>Fresh Diorite</th>
<th>Decomposed Diorite</th>
<th>Calculated</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Percent. Remaining</td>
<td>Per cent. Lost</td>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
<td>46.75</td>
<td>42.44</td>
<td>62.69</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>17.61</td>
<td>25.51</td>
<td>100.00</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>16.79</td>
<td>19.20</td>
<td>78.97</td>
</tr>
<tr>
<td>CaO</td>
<td>9.46</td>
<td>0.37</td>
<td>2.70</td>
</tr>
<tr>
<td>MgO</td>
<td>5.12</td>
<td>0.21</td>
<td>2.83</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.55</td>
<td>0.49</td>
<td>61.26</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.56</td>
<td>0.56</td>
<td>15.13</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.25</td>
<td>0.29</td>
<td>80.11</td>
</tr>
<tr>
<td>Ignition</td>
<td>0.92</td>
<td>10.92</td>
<td>100.00</td>
</tr>
</tbody>
</table>

Data such as those given above should be interpreted with caution since weathering processes vary considerably in intensity with the climate. Further, it is not always easy to be sure that a particular weathered material exactly corresponds with the subjacent rock.

Possible Modes of Formation of Clay

In considering the origin of the clay complex by the chemical weathering of silicate minerals, it is evident that two different types of product are possible. On the one hand, the secondary products which constitute the clay complex may originate from silicate minerals by the removal of certain elements or groups of elements from their crystalline lattices, the removal being accompanied in some cases by replacement with hydrogen or hydroxyl groups. The products may retain the same general lattice form or, more probably, form modified structures.

On the other hand, the origin of the clay complex may consist, as supposed by Van Bemmelen and others, in precipitation from the soluble products of silicate hydrolysis. Whilst, according to the precipitation theory, as originally propounded, the products are absorption compounds of indefinite composition, it is equally possible that the pre-
cipitates may be definite, though complex, in composition. The difficulty in assigning formulae to the clay complex may be due either to the presence of varying proportions of relatively simple crystalline compounds or to the complex being built up as a highly complicated lattice whose structure cannot be expressed by any simple formula.

H. G. Byers\textsuperscript{9} has considered the possibilities involved in the hydrolysis of orthoclase felspar, a representative of the most important group of soil-forming minerals. Assigning to it the simplest of the structural formulæ proposed, the successive steps in its degradation may be pictured as follows:

\[
\begin{align*}
\text{SiO (OK)} & \rightarrow \text{SiO(OH)} & \rightarrow \text{SiO(OH)} \\
\text{Si-O-SiO} & \rightarrow \text{Si-O-SiO(OH)} & \rightarrow \text{SiO(OH)} \\
\text{Al} & \text{Orthoclase} & \text{Al(OH)} & \text{Montmorillonitic acid} & \text{Al(OH)} & \text{Pyrophyllitic acid} \\
\text{Si(OH)}_2 & \rightarrow \text{H}_2\text{SiO}_4 & \text{Al(OH)}_3 & \text{Orthosilicic acid and aluminium hydroxide} \\
\text{O} & \text{Halloysitic acid} & \\
\text{O} & \\
\end{align*}
\]

Montmorillonitic acid is purely hypothetical, the nearest approach to it in nature being the mineral montmorillonite. Pyrophyllitic acid is also hypothetical, differing from pyrophyllite by two molecules of water. Halloysitic acid corresponds with halloysite. It is suggested that pyrophyllite, related to pyrophyllitic acid, is characteristic of podsols and chernozems, due allowance being made for the replacement of aluminium by iron. In the grey-brown podsolic
(brown earth) and ferallitic soils a complex of the halloysitic type may be expected, with admixture of pyrophyllite in the first case and of an unnamed complex $3\text{H}_2\text{O}.\text{Al}_2\text{O}_3.\text{SiO}_2$ in the second.

It is probable that both residual products and precipitation products occur in the complex of ordinary soils, but there are, at present, no methods available for their exact differentiation. It is possible that residual products predominate in the weathering-complex of strongly leached soils, whilst, with incomplete or impeded leaching, and, in illuvial horizons, precipitation products predominate. The precipitation products may include, in addition to constituents deposited from solution at the seat of weathering, illuvial material transported from other horizons of the soil or from the ground-water. These possibilities have been set out on page 44, where the subject is discussed from another aspect.

The course of chemical weathering may be to some extent influenced by the character of the parent rock. Thus J. A. Bonnet\textsuperscript{10} found that the immediate product of the weathering of a basic rock (andesitic tuff) in Puerto Rico was of a sesquioxide character, corresponding with neutral or alkaline conditions. From less basic rocks, e.g. quartz-diorite, the weathering products are less markedly sesquioxide.

F. Hardy and G. Rodrigues\textsuperscript{11} also have reported the formation of a primary sesquioxide weathering product (laterite) from basic igneous rocks. This was also noted by J. B. Harrison\textsuperscript{12}, who considered that percolating waters containing silicic acid in solution could effect a silication of the primary weathering product to give a clay of the kaolinitic (siallitic) type. Hardy also considers that acidic rocks do not give rise directly to lateritic material. Hardy and Rodrigues studied also the weathering of a basic igneous rock, dolerite, in N. Carolina, and found the primary weathering crust to consist of chlorite, secondary quartz, gibbsite, and an unidentified constituent corresponding with the formula $2\text{Fe}_2\text{O}_3.\text{SiO}_2\text{H}_2\text{O}$. 

R. P. Humbert and C. E. Marshall from a detailed study of the weathering of granite and diabase respectively under the humid and warm-temperate conditions of Missouri conclude that the chemical weathering of silicate minerals begins with the interposition of $\text{H}_2\text{O}$—dipoles between silicate anions and their associated cations. The extent of this reaction depends on the solid geometry of the original structures which governs the ease of access of the intruding molecules. This change affects structure and stability to a varying extent, probably only slight in the case of feldspars. In the next stage the modified structure reacts with mobile cations in leaching waters, the principal change being the replacement of the activated cations by hydrogens, which according to E. R. Graham may originate from previously formed acid clays. The final result is the breakdown of the original lattice with formation of secondary clay minerals such as kaolinite and beidellite.

The end point of chemical weathering would be reached when all silicic acid, alkalies, and alkaline earths have been removed. The residual material would then consist of a mixture of hydrated alumina and hydrated ferric oxide, together with original minerals such as quartz that are unaffected by chemical weathering. Material approximating to this constitution may be found in many parts of the tropics and is termed laterite or, preferably, ferrallite.

Under arid conditions, i.e., where percolation is insufficient to remove the products of weathering, hydrolysis may be arrested at a stage in which the weathered product differs but slightly in gross composition from the parent rock. At this stage the alkalies and alkaline earths liberated by hydrolysis may appear in the weathered product as carbonates or sulphates. Instances of weathered materials containing notable proportions of soluble sodium or potassium salts are rare, but it is not uncommon to find in arid climates partially weathered rock material with free calcium carbonate deposited in veins and cracks throughout the mass.

B. B. Polynov distinguishes a number of successive
stages in weathering. He makes a distinction between the residual material or ortho-eluvium and the products of accumulation, i.e., constituents removed by weathering and deposited elsewhere. The stages thus distinguished are:

I. Production of coarse detrital ortho-eluvium as residue, with clastic drifts as products of accumulations.

II. Removal of chlorides and sulphates, which are found in the products of accumulation, leaving calcareous ortho-eluvium as residue.

III. Dealkalization, resulting in siallitic ortho-eluvium as residue, with carbonate-rich products of accumulation.

IV. Desilicification, resulting in allitic ortho-eluvium as residue, with siallitic (allophanic) products of accumulation.

Derived accumulation products may be subjected to further weathering as para-eluvium and neo-eluvium.

These stages of weathering cannot be dissociated from climate. Chemical weathering is intensified by warmth and moisture and is thus most rapid in the humid tropics, where examples of lateritic (ferrallitic) materials corresponding with Polynov's allitic ortho-eluvium are common. The earlier stages are found under less intense conditions of weathering. The question arises whether there is a stage of weathering corresponding with each set of climatic conditions. In Polynov's view the stage of weathering is merely a reflection of the extent to which weathering has proceeded, and if this is so the final stages of weathering should, theoretically, be attainable even under conditions such as aridity and low temperature usually associated with the first stages.

Soil formation, with profile development, may be initiated on any type of ortho-eluvium or accumulation product. For the purpose of soil classification it is desirable to bear strictly in mind the distinction between weathering, which produces the material from which the soil is developed, and pedogenesis, which results in the development of a soil profile.
The degree of weathering is reflected in the character of the different fractions recognized in mechanical analysis. At the one extreme are highly weathered materials of the humid tropics. In such materials, the only original minerals are those that are resistant to hydrolysis, such as quartz, magnetite, etc. Where the parent rock contains no unweatherable minerals, the resulting weathering may consist entirely of the weathering-complex. Some materials on analysis may yield up to 90% of clay. In other cases there is present a certain amount of concretionary material of secondary origin among the coarser fractions. Indeed, in some soils, the greater part of the soil may be represented by concretionary stones and gravel.

The removal of silicic acid, alkalis, and alkaline earths in weathering proceeds to varying stages according to the conditions which obtain. In humid temperate climates there is some evidence that a complex is formed with a molecular ratio of silica to sesquioxides of 2.0. Under humid tropical conditions it appears that de-silicification can proceed further. The weathering products that occur in many parts of the tropics consist, in extreme cases, almost entirely of hydrated sesquioxides and might be described as "dead soils." It is not certain, however, that these are simple residual products. In most cases they are considerably enriched in sesquioxides by precipitation from solutions. Until methods have been perfected for distinguishing between residual products and precipitation products, it is impossible to decide what is the actual end-point of weathering under given conditions.

In the less weathered soils and parent materials of temperate and cool regions there is generally a considerable proportion of weatherable minerals such as felspars, micas, and ferromagnesian minerals in the silt and sand fractions. As a rule, soils derived from unmetamorphosed sediments are more highly weathered, i.e., contain less weatherable material, than soils of primary origin. This is due to the
fact that they have already been submitted to one or more cycles of weathering. It is, indeed, quite possible that, even in Northern Europe, soils and parent materials may occur which consist simply of quartz and the weathering-complex.

An approximate indication of the degree of weathering of a material may be obtained from the proportion of alumina in the clay fraction relative to the total amount of alumina present. Since the principal weatherable silicates are all complex silicates of alumina and other bases, it follows that in a completely weathered material, no alumina will be found in that portion from which the clay or weathering-complex has been removed. Lateritic soils with concretionary alumina in the coarser fractions would, of course, be exceptions to this generalization.

In soils and materials where the weathering has been partly physical, the non-clay fraction contains unweathered aluminosilicate minerals, and the proportion of clay alumina to total alumina may give an approximate measure of the extent to which chemical weathering has proceeded.

Similar reasoning may be applied to the distribution of ferric oxide between clay and non-clay, but its applicability is limited by the possibility of the occurrence of unweatherable iron minerals such as magnetite and by the readiness with which hydrated ferric oxide forms coarse concretions.

**Table VII.—Degree of Weathering of Soils as Shown by Content of Alumina and Ferric Oxide in Clay and Total Soil.**

<table>
<thead>
<tr>
<th>SOIL</th>
<th>Ignited Clay</th>
<th>Per cent. Al₂O₃ in</th>
<th>Degree of Weathering</th>
<th>Per cent. Fe₂O₃ in</th>
<th>Degree of Weathering</th>
</tr>
</thead>
<tbody>
<tr>
<td>Schistose subsoil, Wales</td>
<td>14.0%</td>
<td>4.4</td>
<td>11.3</td>
<td>19.7</td>
<td>39.1</td>
</tr>
<tr>
<td>Shaly soil, Wales</td>
<td>21.0%</td>
<td>7.05</td>
<td>17.5</td>
<td>40.3</td>
<td>2.55</td>
</tr>
<tr>
<td>Shaly sub-soil, Wales</td>
<td>19.5%</td>
<td>8.5</td>
<td>19.7</td>
<td>40.0</td>
<td>2.3</td>
</tr>
<tr>
<td>Red loam, E. Africa</td>
<td>35.5%</td>
<td>13.3</td>
<td>19.0</td>
<td>2.5</td>
<td>8.35</td>
</tr>
<tr>
<td>Yellow soil, India</td>
<td>20.2%</td>
<td>5.35</td>
<td>9.6</td>
<td>3.5</td>
<td>8.65</td>
</tr>
<tr>
<td>Alluvial soil, Holland</td>
<td>48.5%</td>
<td>12.85</td>
<td>13.7</td>
<td>2.95</td>
<td>3.8</td>
</tr>
<tr>
<td>Chalky soil, England</td>
<td>28.5%</td>
<td>4.4</td>
<td>5.0</td>
<td>2.2</td>
<td>6.9</td>
</tr>
<tr>
<td>Red clay loam, E. Africa</td>
<td>50.0%</td>
<td>17.05</td>
<td>19.0</td>
<td>8.25</td>
<td>7.7</td>
</tr>
</tbody>
</table>
The sesquioxides in the clay and in the total material of some typical soils were determined by G. W. Robinson and M. Richardson, and the results are shown in Table VII. The ratio of clay alumina to total alumina is given as a tentative measure of the degree of weathering. The corresponding ratio deduced from the ferric oxide figures is given with more reserve, for the reason indicated above.

The figures show the marked distinction between the relatively juvenile Welsh soils, still in the early stages of chemical weathering, and the alluvial and chalk soils in which, presumably, the material has been submitted to one or more cycles of weathering. The Indian yellow soil and the African red loam occupy an intermediate position, whilst the E. African red clay loam is almost completely weathered. Like the Welsh soils, these are products of primary weathering, but the process has proceeded to a greater degree, particularly in the African red clay loam. It may be presumed that a soil will show a higher degree of weathering than the parent material from which it has developed.

The data for ferric oxide generally confirm those for alumina, but the lower degree of weathering indicated suggests the presence of secondary ferric oxide or unweatherable iron minerals in the non-clay material.

H. Jenny has proposed to use the K/Na ratio in soils as a measure of the degree of weathering. Since sodium is more readily removed than potassium by leaching processes, the tendency is for the K/Na ratio to widen, with maturity of development. The term shifting value, defined as the K/Na ratio in the weathered product compared with the corresponding ratio for the original rock has been proposed.

**BIOLOGICAL WEATHERING**

The presence of vegetation accelerates weathering processes both through the production of carbon dioxide in respiration and also by furnishing material for humus, including humic acids. Two cases, however, can be given in which it may be permissible to speak of biological
weathering. The first case is the action of plant roots in widening cracks and crevices in rocks. This is seen most evidently in the shattering action of the roots of trees growing in thin soils over rock; but it may be supposed that a similar disruptive action is exerted also on a smaller scale wherever plant roots can penetrate the interstices of rocks. The second case is less evident. It is the flaking-off effect produced by the alternate wetting and drying of colloidal material in intimate contact with rock or mineral surfaces. This action may be demonstrated in the laboratory by spreading a layer of starch paste or potato pulp in a Petri dish and drying at 100°. On removal of the dry material it will be seen that the glass has been etched by the flaking off of small laminar fragments. In nature, this action may be produced by lichens and fungi, and, according to M. A. R. Khan, by the roots of certain rock plants.

SOILS OF SECONDARY ORIGIN

In our discussion of the weathering processes we have considered mainly the origin of the weathered material from the direct decomposition of crystalline rocks. Soils derived from such materials may be termed primary soils. Many soils, and amongst them a considerable proportion of English soils, have not originated thus directly, but have been formed from deposits that have already passed through a cycle or cycles of weathering, transport, and consolidation. Such deposits as the Kimmeridge Clay and the Gault Clay have been formed from material that was produced by weathering in past ages, removed by the ordinary processes of denudation, and deposited as sub-aqueous sediments. It is obvious that the weathering-complex of such soils is not to any great extent of contemporary origin. In the absence of heat metamorphism, leading to recrystallization, we may regard all sedimentary rocks as containing the weathering complex formed in the epoch of their deposition. B. Polynov has suggested the descriptive, though cumbersome, terms mono-
chronogenous and heterochronogenous for what we should prefer to term primary and secondary soils, respectively.

Clay fractions of soils derived from unaltered sediments are richer in silica than clay fractions from soils of primary origin. Thus, the average molecular silica-sesquioxide ratio of the clay fraction of a number of N. Welsh soils derived by primary weathering from crystalline rocks was found by G. W. Robinson to be 1.85, whilst the corresponding ratio for a number of soils derived from alluvia and other non-metamorphosed aqueous deposits was 2.67. This may be attributed to the concomitant precipitation of silicic acid from terrigenous waters at the time of their deposition. We have seen that hydrolysis involves a de-silicification of the parent minerals. With free percolation, this excess of silicic acid finds its way into the drainage and river system. It would thus appear that in the weathering-cycle, the silica, separated from its original minerals in hydrolysis, rejoins the weathering products during the stage of sub-aqueous deposition. From data obtained in the writer’s laboratory it appears that the clay complex of calcareous sediments is considerably more siliceous than that of non-calcareous sediments. This is in accord with general experience that a high lime status is conservative of silicic acid in the clay complex.

TRANSPORT OF WEATHERED MATERIAL

In the absence of agencies of transport the products of rock weathering would accumulate in place and eventually form a mantle sufficiently deep to protect the underlying rock from weathering action. The superficial layers of the regolith are continuously undergoing losses or accretions. Broadly speaking, losses occur from uplands with convex relief, and the transported material tends to accumulate in lowlands and regions of concave relief. Characteristic deposits result from the action of the principal agencies of transport.

(1). Gravity. There is a general tendency for loose superficial material to work downhill. This tendency is
SOILS

accentuated where land is stirred by cultivation. It is also favoured by the action of frost, which may result in a downward flow or slip of the surface material known as solifluction. Material accumulated at the foot of slopes is termed hill wash or colluvium. Colluvial accumulations may be seen also on the upper sides of walls and fences running along the contours of hillsides. There is a general tendency for the depth of soil to increase from the top to the bottom of any slope.

(2). Running water. In the sum, the movement of weathered material in rivers is the most important of all agencies of transport. The weathered material from which sedimentary rocks have been built up has, in the great majority of cases, passed through the stage of river transport. Ultimately, water-borne material finds its way to the sea, or in some cases to inland lakes, and is deposited as sub-aqueous sediments. The coarsest material is deposited near the coast line whilst the finer material is carried further from the shore. When consolidated to rocks, such material forms a gradation from conglomerates through grits, sandstones, and mudstones to the finest clay shales. If uplift takes place before the sub-aqueous sediments have been metamorphosed, with consequent recrystallization, they can at once form the parent materials of soils. Where as in the case of many of the more ancient sedimentary rocks metamorphism has resulted in extensive recrystallization, weathering must precede the development of a profile.

In addition to the sub-aqueous sediments deposited near shore lines, material may also be deposited as alluvium along the lower courses of rivers, in deltas, and in estuarine flats.

(3). Moving Ice. Ice is an agent both of transport and weathering. In most parts of northern Europe and the northern parts of N. America may be seen great spreads of material transported by glaciers. This transported material as left by the glaciers is termed boulder clay or till and shows no definite stratification. Associated with the boulder clay or till are stratified sands and gravels representing material
originating from the boulder clay and re-sorted by running water during the period following the recession of the glaciers. Deposits of stratified sand and gravel may also occur in eskers, which were formed by streams running within or under glaciers.

(4). Wind. The most obvious wind-borne deposits are sand dunes, which occur at many points along our coast line, but most commonly where large stretches of sand are uncovered at low tide with land on the leeward side of the prevailing winds. Wind-borne deposits are found also in inland regions particularly in and around deserts. From the soil-forming standpoint, the most important type of wind-borne deposit is the loess, which extends over vast areas of Europe, Asia, and North America. Typically, it consists of a very uniform fine grained material with fine sand or silt as the dominant fraction. It is generally considered to be mainly material transported by wind during and after the Glacial Period. It is, in effect, wind-sorted morainic material.

Apart from deposits that can be certainly identified as wind-borne material, there is in all superficial deposits a certain proportion of material of aeolian origin. This material may represent far-travelled material from deserts or volcanic dust. Even in humid regions, a certain amount of material may become air-borne during dry periods.

BIBLIOGRAPHY


CHAPTER IV

SOIL FORMATION. II. PROFILE DEVELOPMENT

In the preceding chapter we have discussed the processes involved in the formation of the parent material of soils. We must now discuss the development of the soil profile from the parent material. The treatment of these two processes as distinct stages must not be thought to imply that they are necessarily successive in time. The development of the profile may proceed pari passu with the weathering of rock to form parent material. Also, even where the parent material can be definitely distinguished from the soil horizons, weathering still proceeds so long as any weatherable minerals are present.

In the processes of profile development definite horizons are developed. The complete succession of horizons down to and including the upper part of the undifferentiated parent material constitutes the soil profile. The actual soil horizons are generally referred to as the solum. Ideally, the solum represents the parent material as modified by the soil-forming or pedogenic processes.

One of the most difficult problems confronting the pedologist is that of the correspondence of the solum with the underlying parent material. It would greatly facilitate the study of pedogenesis if it could always be assumed that the solum has been developed from material identical with or similar to the parent material. Often, this is clearly not the case, as when sandy over-washes are present. But even where there is no clear evidence that we are not dealing with a vertical change in parent material, proof of uniformity is desirable. C. H. Edelman proposed the use of heavy mineral assemblages as tests of autochthony and this line
of attack has been further developed by C. E. Marshall\(^2\) who used zircon as an assumed immobile indicator and calculated the gains and losses in the different horizons of a profile during its development. The losses from the A and the gains in the B horizon were calculated on this assumption. It was also suggested that the sand fraction 0.125-0.046 mm. was apparently immobile. If this should be confirmed it might prove a convenient starting point for profile studies.

Since the soil profile represents the result of all the soil forming processes, it is the natural unit of study. And therefore, in comparing soils for the purpose of classification, it is necessary to consider, not merely the surface soil, but all the horizons of the soil which constitute the profile. A classification of soils is a classification of soil profiles.

The differentiation into horizons has not always proceeded to its fullest expression possible under the local conditions of soil formation. Profiles which have not attained full development are termed immature or undeveloped. A soil profile may be immature where, as in the case of soils formed from recent alluvial or aeolian deposits, time has not sufficed for mature development. In other cases removal of the superficial material by erosion may have proceeded at such a rate as to keep pace with profile development and maintain a state of immaturity. This is the case on steep slopes under heavy rainfalls. Where, owing to interference with the natural cover of vegetation, erosion has intruded on a mature profile, a truncated profile, characterized by the removal of certain horizons, may result. Examples of truncated profiles are to be found in the South-Eastern United States, where deforestation by early settlers in a region subject to torrential rains and with a rolling topography has resulted in widespread erosion of the surface soil, with the result that over large areas the present cultivated soil represents a former sub-surface horizon.

Finally, even where erosion has not resulted in the prevention or destruction of soil profile development, cultivation may result in the confounding of the natural soil hori-
zons. The typical agricultural soils of Britain thus show disturbed profiles. Apart from the interference with the natural course of profile development occasioned by the processes of cultivation and manuring, and the possibility of erosion after the removal of natural vegetation, the ordinary tillage operations cause a mixing of all the soil down to the limit of their action. And, therefore, if any vestiges of the original profile are to be found, they will occur below the depth affected by ordinary operations, i.e., about nine or ten inches.

Mature profiles are to be found, if at all, in areas of fairly level or gently rolling topography under natural vegetation which have been for long ages subjected to the pedogenic processes characteristic of the locality. In some cases it may happen that profile development, interrupted by erosion, may be resumed on the establishment of a cover of natural vegetation. This leads to the formation of secondary profiles, such as may be observed in some of the uplands of Wales, where grassland or heath has been established in the place of former forest.

Although catastrophic erosion results in the formation of truncated profiles normal erosion must be considered as a factor in profile development. In normal erosion there is a slow removal of surface material and a corresponding deposition of the material thus removed at other sites. In a hilly country like Wales, sites with convex relief are losing material, whilst sites with concave relief show accumulation. This is reflected in the depth of the solum. There is also a normal transport and accumulation through the action of wind, particularly under dry climates.

In considering the processes of profile development it will be convenient to consider them under two headings, namely, (1) the organic matter profile and (2) the mineral profile. This is merely for convenience of treatment, for the processes resulting in the development of the organic matter profile are not completely distinct from those resulting in the mineral profile; indeed the second group of processes
affects the distribution of organic as well as mineral matter. Likewise, the organic profile determines to a large extent the character of the mineral profile. Since the first stages of soil-development result from the addition of organic matter from the residues of vegetation, they must be considered first.

**ORGANIC MATTER PROFILE**

The first stages of soil development may be seen wherever vegetation establishes itself on a mineral substrate. These stages may be seen where lichens and mosses invade the crevices of bare rock or where a weed flora gradually invades raw material such as that presented by the slopes of railway cuttings or the spoil heaps from mine workings. Over most of Britain, this process began with the close of the last glacial period something over 20,000 years ago, when the spreads of vegetation-free material left after the melting of the glaciers began to be colonized from the surrounding non-glaciated regions to the south.

We shall consider in more detail in a later chapter the processes whereby the residues of plant growth are changed through the work of the micro-organisms of the soil into the characteristic organic matter of the soil, namely, humus. In the present chapter we shall consider merely the building up of the organic matter content of the soil and its distribution and arrangement in the soil profile. Organic matter is added to the soil in two ways, namely, from the surface in the form of leaf fall or the dying back of seasonal growth and within the soil from remains of roots. The addition of organic matter to the soil is much less effective by way of surface residues such as leafy fall than by way of the roots of vegetation. Indeed, under close-canopy forest, the content of organic matter below the top few inches is comparatively small. On the other hand soils under gramineous cover, either natural as in downland or prairie, or artificial as in man-made grassland, are always higher in organic matter than in adjacent forest soils.

The organic matter status of the soil profile depends in general on the balance between accessions from plant residues
and oxidative decomposition by micro-organisms. Organic matter status may be low even under a luxuriant cover of vegetation if, as under tropical conditions, biological decomposition is rapid. On the other hand, under the cold humid conditions of a northern moorland, the scanty annual plant growth may be almost completely humified to form peat. Each particular site has its own set of conditions. Starting from bare material, the organic matter status will build up pari passu with the development of the vegetation and presumably reaches an equilibrium condition when the climax stage of vegetation is reached. Any disturbance through human interference must alter most of the factors and a gradual adjustment to a new equilibrium sets in.

H. Jenny has studied the relationships between organic matter content and different soil-forming factors. Under uniform moisture conditions, there is a general tendency for organic matter and total nitrogen to decrease with increasing temperature. From available data a relationship of the type

\[ N = C e^{-kT} \]

is suggested, when \( N \) is the total nitrogen or organic matter, \( T \) the temperature, and \( C \) and \( K \) constants.

The relationship between organic matter and rainfall is complicated by the effect of temperature. With constant temperature and the same type of rainfall distribution there is a strong correlation between rainfall and organic matter content. H. Jenny and C. D. Leonard, from a study of soil samples along the 11 on the loess belt in the United States, found a relationship between percentage \( R \), which could be expressed by the equation:

\[ N = 0.00655R - 0.023 \]

The relationship may be thrown into a more general form by expressing the humidity as a moisture index.

S. Hénin and M. Dupuis have attempted to trace the variation in organic matter content, assuming constant
annual additions of organic matter. This can be expressed by the equation:

$$dy = (K_1 \omega - K_2 y)dt$$

where $y$ is the percentage of organic matter in the soil, $\omega$ is the percentage annual addition of organic matter and $K_1$ and $K_2$ constants. $K_1$, termed the isohumic coefficient, depends on the type of material added whilst $K_2$ expresses the rate of destruction under the conditions prevailing. At equilibrium $K_1 \omega = K_2 y$ and $y = \frac{K_1}{K_2} \omega$. With an initial organic matter content of $Y$ and assuming no additions, the organic matter content $y$ after time $t$ is given by the equation $\frac{Y - y}{Y} = K_2 t$. The value of $K_2$ under conditions of southern England and northern France is of the order 0.015-0.012.

Apart from the general organic matter status of the soil profile, the actual distribution will be affected by the moisture-air conditions within the profile. The distribution of organic matter below the surface layer follows roughly the root development. Where there is an impervious or water-logged layer roots cannot function, and this condition is reflected in a sharp falling off in the organic matter content. On the other hand in well-drained soils the organic matter, as is shown by dark humic colour and as can be confirmed by laboratory analysis, gradually decreases from the surface downwards. Another important factor affecting the distribution of organic matter in the soil is the activity of the soil fauna, principally earthworms in temperate climates, and termites in the tropics.

The organic matter of the soil itself is present principally as complex structureless humus. Humus is intimately associated with the clay constituent of the soil to form the so-called clay-humus complex, but part of it may occur independently. In some soils it is possible to distinguish a dark-coloured coating, presumably of humus, on the surface of the larger particles and structural aggregates. Under
certain conditions humus may become mobile in colloidal solution and undergo deposition in a sub-surface horizon. This is seen in podsol profiles in the so-called $B_1$ horizon, sharply distinguished from the overlying bleached $A_2$ horizon from which humus has been leached out.

**SUPERFICIAL ORGANIC HORIZONS**

Under natural conditions the residues of vegetation and the products of their decomposition may accumulate at the surface and form the upper horizons of the soil profile. Considerable attention has been given to the nomenclature and classification of forest humus, but the problem applies to the soils of all types of natural vegetation, and is not without significance in artificial grassland soils. In the conventional nomenclature used for profile description, the sign $A_{oo}$ is used for undecomposed plant residues such as recently fallen leaves, $A_o$ for the more or less humified organic layer, and $A_i$ for the upper portion of the mineral profile darkened by humus infiltration. The $A_o$ layer is sometimes subdivided into the $F$ layer of decomposing litter and the $H$ layer in which the material is humified.

P. E. Müller⁶, in his classical work on the forms of humus, distinguished between neutral humus or "mull" and acid humus or "torf"; with "torf" Müller included the so-called *raw humus*. Mull, according to Müller, comprised the crumbly loosely stratified material mainly resulting from the activity of the soil fauna. He distinguished "earthworm mull" and "insect mull," the former rich in mineral matter and the latter mainly organic matter.

Numerous classifications of forest humus have been proposed. That of H. Hesselman⁷ is an example of the modern approach to the problem. He distinguishes:

1. Mull, a humus layer with marked crumb structure and containing mineral matter.
2. Mar (or Mor), similar to raw humus but looser.
3. Raw humus, a compact humus layer knit together with fungal hyphae and higher plants.
A classification that has been accepted by many soil workers in the United States is that of S. O. Heiberg and R. F. Chandler. They make a fundamental distinction between mull and mor, mull having its generally accepted meaning whilst mor corresponds generally with the duff, mor, moder, and raw humus of other writers. Mull is organic matter mixed with mineral matter in compact or crumb structure whilst mor is practically pure organic matter more or less compacted or felted together. They distinguish five sub-varieties of mull and five sub-varieties of mor, the distinctions depending on structure and arrangement. More recently W. Laatsch has made the distinction between deposite humus (Auflagehumus) and soil humus. Auflagehumus can be subdivided into "moder" which is finely divided loose humus; raw humus, which is of a fibrous structure owing to retarded decomposition; and "Auflagetorf," which is closely felted raw humus with peaty structure. The term soil humus as used by Laatsch means humus deposited with the soil itself, for which the term "Dauerhumus" is proposed. This is presumably structureless.

F. Hartmann has recently proposed the recognition of the following categories of forest humus:

I. Mull.
II. Moder (Heiberg and Chandler's mor).
III. Surface peat.
IV. Gyttje.
V. Humus sols.
VI. Humus gels and their dehydration products.

Mull is considered to be mainly formed by the excreta of the soil fauna, the coarser varieties being the work of worms, and the finer varieties the work of insects, millipedes, etc. Depending on the character of the mineral substrate he distinguishes calcareous mull, saline mull, mild mull, and acid mull.

Moder, corresponding generally with the mor and raw humus of other writers, is considered by Hartmann as material whose biological decomposition has been arrested,
still showing cellular structure, sometimes more or less felted together with fungal hyphae. When it lies directly on the mineral soil it is termed "deposit moder" (Auglagemoder), within the soil it is termed "soil moder" (Bodenmoder).

Surface peat is distinguished from the true peat of deep deposits in that it is still the seat of biological activity. He distinguishes dry peat, exemplified by the peat of dry heaths, and moist peat developed under continuously moist conditions.

"Gyttje" is a type of humus consisting mainly of animal excreta and deposited in water. As such it can scarcely be reckoned with the other types of humus, which are superficial.

The study of humus types is of special importance for forestry for a close correlation can be observed between the type of surface humus and the character of the site for tree growth. In general the mull types are of higher fertility than the moder or mor types, yet with some species of trees natural regeneration may be better on moder than on mull. Whilst, in general, mull shows a higher base status than moder, this is not invariably the case, for some types of mull may be extremely low in bases.

MINERAL PROFILE

The development of the mineral profile is mainly the consequence of movements of water in the soil, and we may distinguish three possibilities. Firstly, under humid conditions, there is an excess of rainfall over evaporation. There is thus a general tendency to downward movement of soil moisture, and the soil is subjected to a leaching process, whereby certain constituents are carried downwards and either deposited in lower horizons or completely removed in drainage. Secondly, under arid conditions, with an excess of potential evaporation over rainfall, such rain as falls moistens the soil to limited depths. After the cessation of rain, the soil moisture rises again to the surface under the influence of evaporation, with the result that translocation
occurs alternately in both directions without any complete removal of constituents through leaching into the drainage. It should be added that evaporation is not necessarily confined to the actual surface, for when the moisture content is reduced to a certain point, capillary rise is inappreciable. And thus, in the final stages of desiccation, the deposition of materials from solution may occur throughout an appreciable depth of the surface soil. Thirdly, downward movement may be prevented by the presence of ground-water or by the occurrence of an impervious subsoil layer. In such cases, water movement can only occur laterally over the horizon of impedance. The impedance will not always be complete, and intermediate stages can be observed.

The three cases mentioned must not be regarded as hard and fast distinctions, and need some qualification. Thus, in the first place, particularly under monsoon climates, the profile is completely leached during the wet season, but incompletely leached during the dry season. Even in Britain, during summer, particularly in the south-east, the rainfall moistens the profile to a limited depth only, without furnishing drainage. Secondly, even in an arid climate there may be abnormally wet years in which the profile is completely leached with consequent removal of soluble salts to the deeper horizons of the parent material or even to the ground-water level. Finally, in the third category, all degrees of impedance may be distinguished varying from slight impedance to complete obstruction of downward leaching by a high water-table or an impervious stratum.

ELUVIATION AND ILLUVIATION

We may refer to the translocation of material, either mechanically or in solution, as eluviation, and two main types of eluviation may be distinguished, namely, (a) mechanical eluviation, in which, apart from any chemical differentiation, the finer fractions of the mineral portion of the soil are washed down to lower levels, and (b) chemical eluviation in which decomposition occurs and certain products thus
liberated are translocated in true or colloidal solution, to be deposited in other horizons.

It is customary to refer to those horizons from which material has been removed by eluviation as the A or eluviated horizons, and to those horizons enriched by the deposition or precipitation of material from the eluviated horizons as the illuvial or B horizons. The undifferentiated parent material lying below is termed the C horizon. Some writers have applied the term D horizon to the unaltered rock from which the C horizon has developed by weathering.

Mechanical eluviation results in the development of a texture profile, characterized by a light-textured A horizon from which clay and the finer fractions have been removed, and an underlying heavy-textured B horizon enriched by the finer material eluviated from the A horizon. Strongly developed texture profiles are not common in England, but the increase in clay content in passing from soil to subsoil can generally be observed. In the South-Eastern United States, well developed texture profiles are encountered. These show a sandy A horizon overlying a compact B horizon, the so-called hardpan. It may be presumed that the presence of a considerable proportion of sand in the soil is a necessary condition for the development of a good texture profile, for a soil consisting mainly of fine material would not offer the necessary facility for water movements. The lack of strongly developed texture profiles in Britain is mainly due to the relative juvenility of the soils.

The intensity of mechanical eluviation depends mainly on the rainfall; but it is also affected by soil structure. A soil which maintains a well developed crumb structure is less subject to eluviation than a soil in which the single-grain structure prevails. The character of the clay complex and the base-status of the soil are of significance in this connexion.

The formation of hardpans in mechanical eluviation has been studied by I. C. Brown, T. D. Rice, and H. G. Byers, and by H. Jenny and G. D. Smith. The process involves
two stages, namely, the dispersion of colloidal material in the A horizon and its deposition in the B horizon. Two types of deposition are suggested, namely, a "sieve" action in which coagulation by electrolytes plays a part, and attraction based on the adherence of positive sols to quartz grains.

Hardpans formed by mechanical eluviation and deposition must be distinguished from another type of pan, namely claypan, which, according to C. C. Nikiforoff and M. Drosdoff, are formed by weathering in situ of silt particles. Whilst hardpans are irreversibly indurated, there is no permanent induration in the case of claypans. Profiles with developed hardpans are termed planosols by American writers.

Many investigators attach considerable importance to the development of a texture profile as an indication of maturity of development. Whilst this may be valid in comparing soils of the same general constitution, the criterion should be used with caution where soils of different constitution are being compared. Soils with a siliceous clay disperse more readily than soils whose clay is markedly sesquioxidic. We may therefore expect a texture profile to develop less readily in soils of the latter type, where the absence of a texture profile should not be accepted as conclusive evidence of juvenility. Many mature soils in the tropics show comparatively little mechanical eluviation.

CHEMICAL ELUVIATION

Whilst mechanical eluviation is governed chiefly by the intensity of the rainfall and the texture of the soil on which it falls, chemical eluviation involves more factors, and depends on the character of the parent material, the balance of rainfall and evaporation, the nature of the vegetation, and the temperature. The following are the principal groups of constituents affected by chemical eluviation, and the distinctive characters of soil profiles result from their differential movements in true or colloidal solution:
SOIL FORMATION: II. PROFILE DEVELOPMENT

(a) Organic matter.
(b) Silicic acid.
(c) Hydrated alumina and hydrated ferric oxide.
(d) Exchangeable bases associated with clay and humus.
(e) Relatively insoluble salts, principally calcium sulphate and calcium carbonate.
(f) Soluble salts, such as sodium sulphate, sodium chloride, etc.

The simplest type of chemical eluviation is that affecting readily soluble salts such as sodium chloride and sodium sulphate. Under humid conditions with an excess of rainfall over evaporation these salts if formed by the processes of weathering are completely removed from the profile and pass into the drainage. Even under semi-arid conditions they tend to be removed because, although in normal seasons rainfall is insufficient for their removal abnormal seasons occurring from time to time may result in complete leaching. Where, however, the rainfall even in the wettest seasons is insufficient or where percolation is impeded by ground-water or an impervious subsoil, sodium salts may determine the character of the soil profile and give rise to a distinctive group of soils, to be described in a later chapter.

Next in mobility to sodium salts are calcium sulphate and calcium carbonate. Calcium sulphate, although more soluble than calcium carbonate, is less soluble than the sodium salts and may therefore occur in any profile where, under normal conditions, leaching is incomplete. It may occur also in the profiles with impeded leaching. It is only a fugitive constituent in leached soils. Although calcium carbonate is much less soluble in water than calcium sulphate, it has yet an appreciable solubility in percolating soil moisture containing, as it does, carbon di-oxide in solution. Completely leached profiles, if mature, do not therefore contain calcium carbonate. In the incompletely leached profiles of semi-arid and arid climates only the superficial horizons are leached free of calcium carbonate, which accumulates by
precipitation in sub-surface horizons. The amount of this deposition will depend on the lime status of the parent material, and the depth at which it occurs will depend on the effective rainfall. The greater the rainfall, the deeper and less developed will be the horizon of calcium carbonate accumulation. As in the case of calcium sulphate, impedance of drainage will favour deposition of calcium carbonate in the soil profile, so that even in humid regions where freely drained profiles are normally free of carbonates, impeded profiles may show deposition of calcium carbonate. An example of this is the Transvaal Black Turf soils with impeded drainage and a marked deposition of calcium carbonate in the sub-surface horizons. The freely drained Transvaal Red Loams under exactly similar climatic and topographical conditions are completely leached of calcium carbonate.

The presence of an horizon of secondary calcium carbonate in a freely drained profile is the mark of the great world group of soils, sometimes called pedocals, to which the chernozems or black earths and these related groups belong. They are associated with steppe or desert vegetation and climates in which rainfall is less than potential evaporation.

When leaching becomes sufficiently pronounced to affect the whole profile, calcium carbonate ceases to be present as a constant constituent. If profiles are traced across a continent such as the United States or Russia in the direction of increasing humidity, it will be seen that the horizon of calcium carbonate accumulation sinks deeper in the profile and becomes more attenuated, until finally it disappears altogether when the humidity is sufficient for complete leaching. At this critical limit, we need not assume that there is complete leaching of the profile each year. With the moderate degree of leaching implied by these conditions the soil profile, although free from calcium carbonate, has a high base-status and nearly neutral reaction. The potential acidity of the clay and humus is largely neutralized by bases, principally calcium. Under these conditions, the decom-
position of plant residues gives a dark stable humus whilst the clay formed by chemical weathering is of a siliceous type. The structural aggregates or crumbs are very stable and mechanical eluviation is comparatively slight. These circumstances are all favourable to crop growth and to the working of the soil in cultivation. The prairie soils of the United States exemplify these conditions. Well managed grassland soils in Britain, although the conditions are artificial, are essentially similar in their general character.

In the next more advanced stage of leaching the profile becomes more acid through loss of bases. One consequence is that the structural aggregates tend to lose their stability, whereby mechanical eluviation is facilitated. The other important effect of the lowering of base-status is that the clay complex, according to A. Stebbutt\textsuperscript{14}, undergoes a partial decomposition whereby sesquioxides, principally hydrated ferric oxide, are liberated giving yellowish, reddish, or brownish colours to some extent masked in the surface soil by humus. The liberation of sesquioxides may be explained alternatively as the consequence of weathering of primary minerals under the acid conditions prevailing in the solum. K. Lundblad\textsuperscript{15} has shown that certain soils of this general category contain sesquioxides and silica in the form of complexes extractable by the method of O. Tamm\textsuperscript{16}. Whatever be the true explanation, it is certain that soils with reddish, brownish, or yellowish colours in the solum are almost invariably found where the reaction is moderately acid and the leaching unimpeded. In this category are the "brown earths" of temperate climates and the "red loams" of the tropics. They occur generally under forest vegetation, either deciduous or sclerophyllous.

In the types of chemical eluviation we have considered so far, humus plays a subsidiary role. We now have to consider the type of eluviation that occurs where the residues of the natural vegetation accumulate as a partially decomposed superficial layer of raw humus such as occurs under coniferous or heath vegetation. The successive types of eluvia-
tion considered above are marked by progressive lowering of base status and the type of eluviation found under raw humus is commonly considered to be marked by an even more marked degree of acidity. Whilst this generalization may appear valid from a consideration of the examples presented to ordinary experience, we cannot exclude the possibility that raw humus may be formed under neutral or even alkaline conditions.

Given the presence of raw humus as the most superficial layer of the soil profile we find a type of eluviation commonly known as *podsolization*, which results in the most extreme cases in the development of the podsol profiles. Podsols and their variants will be discussed in a later chapter, but in order to understand podsolization as a soil-forming process it will be convenient to give in outline the essential features of a podsol. The term is derived from a Russian folk name for a grey or ash-coloured soil. Podsols occur typically in cool, humid climates under coniferous forest or heath vegetation. The succession of horizons is shown diagrammatically in Plate II. Using the generally accepted nomenclature they are as follows:

A<sub>00</sub>. Leaf litter.
A<sub>0</sub>. Raw humus.
A<sub>1</sub>. Dark coloured humus layer.
A<sub>2</sub>. Ashen-grey or light grey layer leached of humus and sesquioxides, with superficially bleached sand, gravel, and stones. Loose structure.
B<sub>1</sub>. Black or coffee-brown indurated humus layer.
C. Parent material.

We have seen that with a moderate degree of base-unsaturation free sesquioxides become liberated either by decomposition of the clay complex or as a direct consequence of weathering in the solum. In the podsolic type of eluviation, the sesquioxides and humus become mobile and are
Diagrammatic Representation of a Podsol Profile.

- **Vegetation**
- **Raw Humus layer (A₀)**
- **Humus layer (A₁)**

- **Loose structure with eluviation of clay.**
- **Compact structure with illuviation of clay. Greatest compaction in B₁ layer.**

- **Ashen-grey or light grey layer (A₂), leached of humus and sesquioxides. Bleached sand grains, stones, and gravel.**

- **Black to dark brown layer of humus accumulation (B₁). Black staining on stones and gravel.**

- **Brown to orange-brown layer of sesquioxide accumulation (B₂). Brown iron staining on stones and gravel.**

- **Grey or brownish-grey parent material (C).**
leached out of the upper horizons giving rise to a bleached loose structureless layer. Below this bleached layer deposition takes place, first of humus and below this sesquioxides. The deposition of humus is shown by the dark colour and is accompanied by induration sometimes to almost rock-like consistency, giving the so-called ortstein. The presence of considerable proportions of stones or gravel appears to favour this process. The horizons of deposition of the sesquioxides are marked by the brownish or orange colour of hydrated ferric oxide and a certain degree of induration. In certain podsols the leached humus is not precipitated in a definite B, horizon, and there is a direct transition from the bleached horizon (A₂) to the brown or orange coloured B₂ horizon.

Owing to the breakdown of crumb structure in the A horizons there is a certain amount of mechanical eluviation and the clay content may increase markedly in passing from the A₂ to the B horizons.

The mechanism of podsolization is not fully explained. It is generally agreed that the sesquioxides are peptized by humus in solution. Whilst this peptization is generally associated with acidity, it is not necessarily a hydrogen-ion effect. More probably the peptizing effect is due to tannin, anions of humic acids, or anions of hydroxy-acids formed in the decomposition of plant residues.

H. T. Jones and J. S. Wilcox¹⁷ from laboratory studies concluded that hydroxy-acids are effective in dissolving sesquioxides in complex anions, which are mobile and from which sesquioxides are precipitated as basic salts.

A. Demolon and E. Bastisse¹⁸ have shown that in the presence of phosphate, silicate, and humate anions, iron can become mobile, even in alkaline soils. In podsolization iron and aluminium are leached out as colloidal complexes.

P. H. Gallagher¹⁹ assigns a minor role to humic acids and considers that podsolization may be due to the action of comparatively simple acids, such as oxalic acid liberated in the decomposition of organic residues. Humus-sesquioxide
complexes may be formed by the sorption of mobile humus by precipitated sesquioxides. A. A. Rode also suggests that the organic constituents mobilizing sesquioxides are products of incomplete breakdown of plant residues.

The mechanism whereby humus and sesquioxides are deposited in the B horizons is even less clear than the mechanism of eluviation. The simplest explanation appears to be that since podsolic eluviation is usually the accompaniment of extremely acid conditions, the precipitation in lower horizons is due to their higher base status. It is also possible that sesquioxides may be precipitated through restriction of percolation in an horizon enriched in clay by mechanical eluviation. C. G. T. Morison and D. B. Sothers consider that ferric oxide in illuvial horizons arises from irreversible precipitation from humus-protected soils during periods of drought. During winter, movements of soil moisture are mainly downwards and eluviation predominates; with the drying out of the profile in summer, precipitation occurs. The localization of precipitation in a definite horizon is probably connected with the nature of the moisture gradient. If the views of Morison and Sothers are correct, podsol profiles should be shallower in extremely wet climates owing to the smaller depth to which the profile dries out. There is some evidence that this is the case.

Another possibility is that the original tendency to deposition of sesquioxides is given by a restriction of freedom of drainage by the mechanical deposition of fine material in the B horizon. Whilst electronegative soils can pass fine capillaries, electropositive soils such as those of ferric oxide and alumina are precipitated. A defect of this theory is that soil soils appear to be invariably electronegative.

S. Mattson has discussed the processes of eluviation and illuviation from the standpoint of his theories of isoelectric weathering. He divides soils according to their base status into three main groups, namely:

A. "Supersaturated" or alkaline soils.
B. Partly saturated or semi-saturated soils.
C. Negatively base-saturated or acid-saturated soils.
Considering a neutral soil as being 100% saturated, a soil containing free carbonates and alkaline in reaction is supersaturated. As the base saturation falls below 100% we have the region of the partly saturated soils. Negatively saturated soils, according to Mattson, are soils in which combined acids predominate over combined bases. This state of affairs may exist in the sesquioxidic B horizons of podsol soils and in ferrallitic soils—the laterites of some writers. Such soils on electrodialysis yield small amounts of cations but appreciable amounts of anions, principally sulphate.

Soils having a positive base status, i.e., soils of groups A and B, undergo anionic solvation, i.e., loss of anions, mainly silicic acid, leaving a residual complex richer in basoids, i.e., in sesquioxides. Owing to the normal increase in pH in descending the profile no precipitation occurs at lower levels. Where weathering is intense and there is but a slight accumulation of humus acidoids, a relatively high base status will obtain and there will be extensive solvation and eluviation of silica, leaving a residual complex whose ferrallitic character will become more pronounced in passing from temperate to tropical conditions, the final end point being a residual complex consisting of hydrated sesquioxides.

Under cooler conditions, favouring the accumulation of humus acidoids, the prevailing pH will be lower and the residual complex at zero base-status will be of a siallitic-ferrallitic type. The type of humus formed under conditions of high base status readily undergoes anionic solvation and some iron and aluminium is mobilized with it. This, according to Mattson, gives rise to a kind of pseudo-podsolization, distinct from that occurring under raw humus. To this type of eluviation must be assigned the apparent podsolization of solot soils at pH values above those usually associated with cationic solvation. The brown colour of unsaturated siallitic soils, e.g., the brown earths, is attributed by Mattson to oxidation of humus in humic-ferric oxide complexes.

When a soil is at zero saturation solvation is at minimum and in the absence of organic and mineral acids, differential
leaching of acidoids and basoids is absent. But an accumulation of organic matter as raw humus introduces a new factor. The gel complex now undergoes acid hydrolysis and cationic solvation occurs, whereby sesquioxides are eluviated. Aluminium is a stronger base than iron and its compounds have a higher isoelectric point. Alumina therefore undergoes eluviation at a higher pH than ferric oxide. But the presence of ferrous compounds in the complex introduces a complication, for ferrous hydroxide is more strongly basic than alumina. Mattson found ferrous iron in all artificially prepared cationic solvates from podsol profiles. The cationic sol complex as it moves down the profile encounters a higher pH and, unlike the anionic complexes, undergoes precipitation at lower levels. The material in the upper B horizon is richer in humus, and iron oxide is precipitated before alumina. In the early stages of cationic solvation the parent gel is relatively rich in sesquioxides and the resultant sol will therefore yield a gel with higher isoelectric point than in the later stages. It will therefore be precipitated deeper in the profile than will be the case with the less sesquioxidic sol produced in the later stages. The B horizon will therefore tend to develop from below upwards as has been postulated on other grounds by V. T. Aaltonen\textsuperscript{23}.

Eluviation does not consist necessarily in the downward translocation of material. Indeed, even where the net movement is in a downward direction, as in the podsols, the actual deposition in the B horizon may take place from ascending solutions during periods of drought—a type of illuvial deposition which is even more pronounced in certain ferralitic profiles. In those portions of the humid tropics that are distinguished by well marked wet and dry periods, mature soil profiles often contain concretionary material irreversibly deposited during dry periods from relatively concentrated soil solutions. This concretionary material varies in development from small scattered pisoliths to massive ferruginous crusts.

According to H. Harrassowitz\textsuperscript{24}, sesquioxides, under
alternating conditions of humidity, may be carried to the surface as humus-protected soils and there undergo irreversible precipitation. A. Reifenberg\textsuperscript{25} postulates a similar process with colloidal silicic acid as the protective colloid. Under such conditions, superficial deposits of sesquioxides may be regarded as illuvial horizons.

PROFILE DEVELOPMENT UNDER IMPEDED DRAINAGE OR SUBMERGED CONDITIONS

It frequently happens that, either through the presence of a high water-table or through the occurrence of impervious strata, certain soil horizons are submitted permanently or intermittently to complete saturation. Such conditions result in changes which, from the frequency of their operation, must be reckoned among the most important of the pedogenic processes.

The most notable consequence of excessive moisture is the exclusion of air and the restriction of free percolation. The first result of the exclusion of air is the limitation of root development to the immediate surface and the encouragement of a type of natural vegetation adapted to these conditions. In this way, marsh, swamp, or bog associations are developed. Where the soil is continuously waterlogged to the surface, peat develops by anaerobic humification; in the early stages under the influences of ground-water, but later, where the rainfall is sufficient, under the influence of atmospheric moisture alone. Even in cases where the ground-water level is not sufficiently near the surface for peat development, there is generally a development of peaty organic matter, and a sharp change in organic matter content is noted in passing from the humic layer to the horizon in which root development is inhibited by lack of air.

The second consequence of exclusion of air through waterlogging is the prevalence of processes of reduction. Ferric compounds are reduced to ferrous compounds, as shown by the grey or greenish-grey mineral colour of waterlogged horizons and the occurrence of such compounds as
ferrous sulphide, iron pyrites, marcasite, and vivianite. Where there is a fluctuation of the ground-water level the alternation of oxidizing and reducing conditions leads to a characteristic rusty-brown or brownish-yellow mottling. In extreme cases notable deposits of hydrated ferric oxide may be formed, represented by the bog iron ore of northern countries and the "murrum" of the tropics. Manganese dioxide is frequently deposited in horizons subjected to drainage impedance.

The horizon characterized by deposition of secondary hydrated ferric oxide is termed the G or gley horizon. The term "gleying" is, however, not always used in a very precise sense. In some cases, there may be little iron deposition, but impedance is indicated by the greenish grey or bluish grey colour and a characteristic glaze or varnish on the surface of the structural elements. Some would regard this as the characteristic mark of gleying.

The deposition of secondary iron oxides may take different forms. It is most strongly marked where there is a fluctuating water-table in a fairly permeable profile as in many sandy soils in low-lying flats. In heavier soils where the impedance is due to the character of the material and not to a regional water-table, the rusty markings are often less pronounced and more diffuse. In such cases there is the possibility that iron compounds rendered soluble by reduction may have been removed by lateral leaching. The ochre of ditches may originate in this way.

Many grassland soils with impeded drainage show well-marked deposition of rusty material along root channels, an appearance that can serve as a trustworthy sign of seasonal wetness.

The restriction of percolation hinders the removal of products of silicate weathering. An important consequence of this is that the weathering-complex formed under such conditions is markedly more siliceous than that formed under conditions where complete leaching is possible. This, in an equal measure with reducing conditions, contributes to
the production of soils with grey mineral colours, as distinct from the reddish or brownish colours of soils which result when the removal of excess silicic acid is unchecked.

In basin shaped areas where waterlogging is accompanied by high evaporation, soluble salts, including calcium carbonate, gypsum, and sodium salts, may accumulate, giving rise in some cases to gypseous or calcareous deposits or concretions, and in other cases producing saline or alkaline soils.

The incomplete removal of soluble materials, except in the more humid regions, results in a relatively high base status. Under such conditions the humification of organic residues appears to produce a dark type of humus. Such soils as the black turf soils of S. Africa may have dark grey or black colours and yet contain less than 1% of organic carbon. The black colour of certain prairie soils in N. America developed in association with a high water-table or impeded drainage may also be noted.

W. O. Robinson found that, in the presence of organic matter, submergence is followed by a rapid increase in the solubility of iron, manganese, calcium and magnesium. Solution is effected by the presence of carbon dioxide from the microbiological decomposition of organic matter, which can take place with considerable rapidity. The iron comes into solution in the ferrous state, probably as ferrous hydrogen carbonate. Where leaching is possible, considerable losses of iron, manganese, calcium, and magnesium may occur. Where organic matter is absent, or present in small amount, the solubilization of iron, manganese, calcium, and magnesium is negligible. Any acidity directly resulting from submergence is attributable to the carbon dioxide formed by decomposition of organic matter. The formation of soluble carbonates must, however, exert an alkalinizing effect. The subsequent leaching of the bases thus mobilized would cause a lowering of the base-status.

Pan formation may occur in water-logged soils in the zone through which the level of the water-table fluctuates.
In certain types of tropical soils, with impeded drainage, the deposition of iron oxides may result in the development of highly indurated concretionary material, known in Africa as "murrum" and in Cuba as "mocarrero".*

The most extreme instance of pan formation is the development of ferruginous and aluminous crusts in the surface horizons of laterite profiles. These will be discussed in detail in a later chapter. It should be noted that laterite crusts are associated with tropical climates in which alternations of rainfall and drought occur. It is doubtful if they are formed by contemporary processes in tropical regions of unbroken humidity. The development of deposits of sesquioxides varies considerably. Where the tendency to concretionary formation is less strongly developed, pisoliths may occur in the upper horizons. These consist mainly of hydrated sesquioxides, but silica may also be present.

In a succession of soils starting with tropical rain forest and passing to less humid regions, concretionary material begins to be observed under monsoon forest and becomes more developed in savannah, reaching its greatest development in certain types of lateritic soils. In the direction of increasing aridity, sesquioxide concretions become less evident and calcareous pans become increasingly developed as steppe and desert are reached.

Calcareous and gypseous pans are essentially formations of semi-arid and arid climates and do not necessarily imply drainage impedance. In the profiles formed under such conditions, calcium carbonate and calcium sulphate are deposited in distinct horizons, which approach the surface more nearly the more arid is the climate. The deposits of calcium carbonate and gypsum in such profiles are generally of an intermittent character and are scarcely to be considered as pans in the strict sense. In desert and semi-desert regions such as the South African Karroo,

*It is worthy of remark that pans have been given names by farmers in many parts of the world. Thus we have foxbench in England; sentur in Wales, ortstein in Germany, oklip in S. Africa, padas in Sumatra, and nasas in Palestine.
massive rock-like deposits of calcium carbonate may occur at or near the surface. Calcareous pans may occur also under conditions of restricted drainage, as in certain soils of the tropics.

**STRUCTURE PROFILE**

In addition to the texture profile and the mineral and organic matter profiles, we must recognize also the structure profile. We have seen that soil occurs naturally in the form of compound aggregates formed by the grouping of the ultimate particles. The character of the pedogenic processes is often reflected in the type of compound aggregate formed. Thus, in a podsol profile, the superficial horizon tends to have a single-grain structure, becoming loose and dusty in drying. The illuvial or B horizon tends to become compacted into a pan. In the chernozems (Chapter XI), the surface has a porous granular structure, whilst the lower horizons may exhibit nut or columnar structures (see pp. 255 et seq.). Prismatic and columnar structures are also shown by many soils with impeded drainage. Finally, there is the formation of pans, which in some cases may attain such a development as to determine the whole character of the profile.

**SOIL EROSION**

By soil erosion is understood the removal of material from the surface of the soil by the agency of running water, wind, or even by gravity. Where the loss of material is approximately balanced by the formation of new soil from the underlying parent material, we may speak of normal erosion. Where for any reason the loss of material predominates over profile development we have catastrophic erosion. Since the surface layers of the soil are richest in plant food and are the feeding ground of plants, catastrophic erosion involves a definite loss of valuable material and, if it becomes sufficiently intense, may lead to the complete destruction of the soil as the seat of plant life. No hard and fast line can be drawn between normal and catastrophic erosion, unless by catastrophic erosion is understood the accelerated erosion which may result through human inter-
ference by the destruction of the natural vegetative cover of grassland or forest. Even under natural vegetation, however, erosion may be sufficiently intense to prevent the development of a mature profile, and to this extent can be regarded as catastrophic. In any case erosion must be considered as a factor in soil profile development.

Under the most stable conditions, there is always a certain loss of material from the soil surface, but this is more or less balanced by the formation of fresh soil through the processes of weathering and the accumulation of organic matter from the residues of vegetation. Such a balance between erosion and soil formation permits the development of a mature soil profile and is realized under a stable closed vegetation with flat or gently rolling topography. Under these conditions wastage is comparatively slight and may amount, when reckoned on the whole land surface, to a few inches per century. A considerable proportion of this wastage represents material lost by solution in percolating waters which find their way into rivers and ultimately to the sea. Erosion of this kind must be considered as beneficial, since it is accompanied by the continual exposure of fresh layers of material to the weathering processes. In its absence, weathering might eventually come to an end-point.

It is obvious that, where considerable erosion occurs, the natural soil profile is destroyed or truncated and, indeed, may never attain full development. In considering the relationships of the soils of a particular region to the great soil groups, it is therefore necessary to make allowance for the possible effects of erosion in the past. This is particularly the case in humid tropical and sub-tropical regions, where, as will be seen, erosion is intense and has led to considerable modifications in profile development.

Considering in the first place erosion by running water, we may distinguish two types, namely, sheet erosion operating over whole areas, and gullying, in which the erosion is confined to comparatively narrow channels. In the first type of erosion, although the results are rarely spectacular,
its operation over large areas leads to enormous losses of material, which either accumulates in valleys or hollows or else is carried away by rivers to the coast. In the second type, the effects of erosion are evident in the deep channels which are cut; but the damage due to the loss of material is less serious than the resultant lowering of the water-table, which leads to excessive drainage, and damage to vegetation through drought. Examples of this kind of erosion are common in S. Africa and in parts of the United States.

The rate of erosion of soil by running water depends principally on four factors, namely, (1) the amount and intensity of the rainfall; (2) the angle of slope; (3) the presence or absence of a vegetative cover, and, if present, its character; and (4) the nature of the soil itself.

The actual amount of annual rainfall is of less importance than its distribution and the extent to which it occurs in torrential downpours. It is only when the soil is unable to dispose of rainfall by percolation that running water capable of producing erosion appears on the surface. Even in the wettest districts of Britain, it is only in exceptional cases that rain falls so quickly that the soil is unable to dispose of it by percolation. The position is completely different under humid tropical and sub-tropical conditions. Here the rain occurs to a large extent in intense falls of comparatively short duration—as much as six inches per hour having been recorded—with the result that running water soon appears on the surface. Similar conditions occur in certain arid regions where the scanty precipitation may occur mainly in the form of occasional torrential downpours. It is a remarkable circumstance that, in arid regions such as Nevada and Southern California, evidences of erosion and transport of material by water action are more obvious than in more humid climates such as that of Britain.

It is evident that the slope of the land has an important bearing on its liability to erosion. Other things being equal, the amount of run-off and therefore erosion is more
intense on steep slopes. R. E. Dickson\textsuperscript{27} obtained the following figures for run-off in inches from a Texas clay soil:

<table>
<thead>
<tr>
<th>Gradient Level</th>
<th>1 in 100</th>
<th>1 in 50</th>
</tr>
</thead>
<tbody>
<tr>
<td>Run-off water</td>
<td>0.62</td>
<td>3.42</td>
</tr>
</tbody>
</table>

H. H. Bennett\textsuperscript{28} reports a 25% increase in erosion for each 1% increase in gradient over 1 in 27 on a loam soil in Missouri.

Even with the comparatively slight erosion to which British soils are subjected, the accumulation of soil above fences running along contours can be generally observed, and is the more evident the steeper the slope.

The importance of a cover of vegetation in checking erosion is obvious. Not only does it protect the soil from the beating action of rain, but the network of roots and stems hinders erosion even when water appears as run-off. Further, the presence of a closed vegetative cover assists in maintaining the top soil in such a physical condition that it can absorb intense downpours without suffering the mechanical puddling that prevents free percolation.

In regions which, from the amount and nature of the rainfall, are liable to erosion, the destruction of the natural vegetation is fraught with serious consequences unless carefully planned measures are taken to minimize the danger. Unfortunately, over large areas of the tropics and sub-tropics, irreparable damage has already taken place, resulting in the complete removal of the surface soil and the exposure of raw subsoils or even bare rock.

The destruction of the natural vegetative cover by grazing animals may expose soil to the influences of erosion. This has occurred in many parts of the South African Veldt\textsuperscript{19}, where the "kraaling" of stock, rendered necessary on account of the menace of jackals, has resulted in intensive trampling, followed by "slooting" (gullying). The consequent lowering of the water-table has intensified the effects of drought and led to a general lowering of stock-carrying capacity.
Soils laid bare by arable cultivation or by the agency of grazing animals are naturally subject to erosion; but important differences may be remarked in their susceptibility to this process. Generally speaking, it may be said that a granular or crumb structure favours resistance to erosion, whilst a deflocculated or single-grain structure renders soil liable to erosion. The presence of a high proportion of organic matter is helpful in diminishing liability to erosion. It is noteworthy that, even on the steepest Welsh hillsides in cultivation, erosion is almost imperceptible, a circumstance which may be partly attributed to the high proportions of organic matter present in the soil and partly to the mineralogical character of the clay fraction.

H. H. Bennett (loc. cit.) has shown that in the South-Eastern United States, Cuba, and Central America, soils fall into two classes with respect to their liability to erosion. On the one hand, soils whose clay constituent is characterized by excess of sesquioxides are markedly resistant to erosion, probably owing to their tendency to assume a crumb structure, which favours the rapid absorption of rainfall. On the other hand are soils whose clay is of a more siliceous type. These are more readily eroded owing to their tendency to deflocculation.

The extent of erosion in the tropics and certain parts of the sub-tropics is scarcely apprehended by those who are only familiar with British conditions. A measure of its intensity is furnished by the content of suspended matter in river waters. Whilst in the rivers of Western Europe this amount is generally less than fifty grams per cubic metre, from 2,000 to 10,000 grams per cubic metre are found in tropical rivers. From South Africa, silt contents of over 2% are reported. V. A. Beckley and W. C. Lowdermilk have directed attention to the grave menace of erosion in Africa and the United States, respectively.

Bennett (loc. cit.) has estimated that, in the Piedmont region of the United States of America, about 65% of the area which has been or is now under cultivation has lost from
4 to 18 inches of soil, and that more than 50% of the alluvial land formerly cultivated in valley bottoms has been covered up by erosional débris. This process dates from the settlement and deforestation of the area by the white settlers. Profiles are frequently truncated by the removal of the A and even part of the B horizon; so that a large proportion of the area at present under cultivation represents what was originally B horizon material. An appreciation of the extent of the erosion is necessary in order to understand the soil profile relationships of this region, where the assumption that the surface soil represents an A horizon may frequently lead the investigator into serious error.

It may be conjectured that erosion, analogous to that which has proceeded in the Eastern United States within the last three centuries, has taken place also in Great Britain, although at an earlier period, and, in view of the smaller prevalence of torrential rain, to a smaller degree. Evidence for this in Wales is furnished by the observations of G. W. Robinson on the changes in composition of the clay fraction in the soil profile. Whilst the general tendency under Welsh climatic conditions is towards a leaching out of sesquioxides from the surface horizons, many of the surface soils show the high proportions of sesquioxides associated with B horizons, and this tendency is most marked in situations where rainfall and topography might be expected to have resulted in heavy erosion. It may be conjectured that considerable erosion took place when the land was first deforested, with consequent removal of material of a relatively siliceous character, such as is now commonly found in hollows and bottoms. With the institution of grass husbandry and the consequent increase in organic matter, erosion became checked and, if we may draw a parallel from Bennett’s observations in America, the sesquioxidic character of the clay probably contributed to stability in this respect. In the higher lands, the grass vegetation gradually degenerated to heath, and secondary podsols were developed in the primitive B horizon.
The evidence from England is less plain, but in many districts, notably in chalk country, stretches of colluvial soils on slopes, and shallow profiles in the uplands suggest past erosion.

The effects of erosion by wind are more plainly seen in the accumulation of transported material than in the actual denudation which accompanies the process. The great wind-borne loess deposits are the result of wind erosion acting in past ages.

Wind erosion, at the present day, is principally active in desert and semi-desert regions. Since only the finest particles are liable to transport by wind it frequently happens that a sorting out occurs whereby fine material is removed and the coarser gravel and stones remain behind as "desert pavement." This effect is only operative at the immediate surface and finer-grained material is found below the level of stones and gravel.

The occasional occurrence of hard layers of calcium carbonate at the surface in parts of the South African Karroo and other deserts may be due to removal of formerly overlying layers by wind action.

PEDOGENIC FACTORS

In the present chapter we have studied the different processes whereby soil profiles are developed from parent material. Considering soil formation in its most general terms we may say that the final result depends on the following factors or groups of factors:

1. Parent material.
2. Climatic factors.
3. Topography.
4. Vegetation.
5. Time.

And we may perhaps add a sixth factor, namely, human interference. This addition might be avoided if instead of the factor vegetation the term biotic factors were substituted.
H. Jenny\textsuperscript{32} has expressed the relationship between soil properties and the pedogenic factors in the general form:

\[ s = \int (c, o, r, p, t, \ldots) \]

where \( s \) is any soil property such as clay content, organic matter content, etc., \( c = \) climate, \( o = \) bio-sequence (vegetation, etc.), \( r = \) topography, \( p = \) parent material, \( t = \) age. He derives also subsidiary equations of the form \( s = f(t), c_1, o, r \), which expresses the fact that if all factors except one are assumed constant a given soil property is a function of the variable factor, in the example given, age.

Carrying his argument further he gives the following expression for the variation of any soil property from \( s_a \) to \( s_b \):

\[ s_b - s_a = \int_c^d \left( \frac{\partial s}{\partial F_1} \right) dF_1 + \int_e^f \left( \frac{\partial s}{\partial F_2} \right) dF_2 + \ldots \text{ etc.} \]

where the factor \( F_1 \) varies from \( c \) to \( d \), the factor \( F_2 \) from \( e \) to \( f \), and so on. If the definite integrals of the factors \( F_2, F_3 \) are small in comparison with those of \( F_1 \), then \( s_b - s_a \) is mainly governed by \( F_1 \). A number of equations for properties may be devised and all these may be regarded as an ensemble \( E(s) \) which is a function of all the factors present.

An early attempt to express the relationship between soil properties and soil forming factors was that of E. M. Crowther\textsuperscript{33}, who showed that with constant rainfall clay silica-alumina ratio increases with temperature. Jenny has carried this line of enquiry still further and has proposed a number of equations, connecting certain soil properties with soil-forming factors. Whilst some correlations can be well expressed in equation form, it is important that the limitations of this method of approach should be recognized. In the first place some of the factors in soil formation are not susceptible of quantitative expression. This applies particularly to biotic factors and to parent material. Even climatic factors, which as meteorological data, appear to lend themselves to mathematical treatment, involve considerable
complications. For example, crude figures for average annual rainfall can be very misleading. Even when temperature effects are allowed for, it is difficult to see how the seasonal variation of rainfall and its variation from year to year, all significant for soil formation, can be adequately allowed for. Jenny has proposed equations connecting such properties as organic matter content, base-status, and clay composition, with rainfall, temperature, and time. These are mainly based on American data. It remains to be seen if these equations are valid for all parts of the world; but in the meantime they may serve to stimulate enquiry.

It must be recognized that the factors involved in soil formation are not all of equal significance. In the first place, we must regard parent material as participating passively in soil formation. Carrying the enquiry a stage further back we may say that the parent material is the result of the weathering processes on the parent rock, which thus becomes the ultimate passive participant in soil formation. To assign a passive rôle to parent material and ultimately to parent rock is not to minimize its importance. In some cases the nature of the parent rock, determining the character of the parent material, can decisively affect the character of the final product.

By contrast with parent material and parent rock the factors involved in climate must be regarded as determining the active agencies in soil formation. Topography enters in a subsidiary rôle as modifying the operation of those agencies dependent on climate. Vegetation plays an active rôle in soil formation but is itself conditioned by soil, climate, topography, and until a climax is reached, by time. The time factor is on a different footing from other factors. In itself, it cannot have any effect, but intervenes as governing the amount or extent of operation of other factors.

The relationship between soil and the various factors may therefore be thus stated. The character of the soil is determined by the operation of processes dependent on climate as modified by topography. These processes act
both directly and, through vegetation and other biotic factors, indirectly on a given parent material, itself the result of the action of weathering processes on a parent rock. The operation of the weathering and of the soil-forming processes takes place in time, which thus affects, although not causally, the final result.

CLIMATE AS A PEDOGENIC FACTOR

The principal effects of climate on soil formation are through rainfall and temperature. Rainfall governs the supply of moisture to the soil and also its aeration. It also determines the character and extent of the leaching to which the profile is submitted. Temperature has a direct effect on soil formation through its effect on the velocity of chemical reactions, which approximately doubles for each 10° C. rise in temperature. It also governs microbiological activity in the soil. Finally, it affects the efficiency of the rainfall, since losses from the soil through evaporation and through transpiration by crops increase with temperature. Thus a mean annual rainfall of 20 inches under British conditions denotes a humid climate, whereas the same rainfall in the tropics would imply arid or semi-arid conditions. Climate also affects soil formation indirectly through the vegetation, which is mainly governed by climatic and soil conditions.

At an earlier point in our consideration of soil formation, we have distinguished between the processes of weathering, whereby parent material is formed from rock, and profile development, whereby the parent material is modified and differentiated to form the soil profile. In considering the effect of climate and other factors in soil formation, we cannot ignore their effect on the processes of weathering. The problem is complicated by the fact that in some cases the processes of weathering are governed by contemporary factors, i.e., the same factors that influence the development of the soil profile, whilst in other cases the weathering that has produced the parent material took place in an earlier epoch.
The first case is exemplified by shallow sedentary soils developed and developing under British conditions. Here, the various factors affect both the formation of the parent material and also the process of profile development. The two effects can hardly be distinguished and it might be permissible to regard the change from parent rock to soil as a single process or group of processes. The second case is exemplified by soils developed from glacial drifts, alluvia, and aeolian deposits where weathering has occurred remotely in place and, it may be, also in time. Here, if we may neglect changes in unweathered constituents, the contemporary process is simply soil formation.

Under certain humid tropical conditions the relationships are somewhat complicated. Here we may find a considerable depth of weathered material overlying the parent rock. This is, of course, the parent material of the soil. In so far as the parent material determines the character of the resultant soil, the climate and other factors operate both in the stage of weathering and in the stage of profile development, and in correlating a certain soil property with a particular factor it is desirable to decide at which stage it is operative. Thus, many tropical soils have a markedly ferrallitic character. This should generally be attributed to the operation of factors during weathering rather than during actual profile development. But some caution is necessary because weathering may have taken place under climatic and hydrologic conditions differing considerably from those obtaining at present.

Speaking generally, we may say that contemporary conditions may be safely correlated with organic matter, including nitrogen, base status, and structure, particularly when considered in their vertical relationships. On the other hand, the effect of parent material, and consequently the factors operative in its formation, is dominant in determining texture, and the chemical and mineralogical character of the clay complex.

Many attempts have been made to express the relative
humidity of a climate by a single factor with the object of classifying climates. The various methods are all based on the relationships between rainfall and evaporation. A humid climate may be defined as one in which rainfall exceeds evaporation and an arid climate one in which evaporation exceeds rainfall. The difficulty in applying this definition is that of obtaining an accurate measure of evaporation. Since evaporation is in a large measure controlled by temperature, numerous methods have been proposed for expressing humidity as a function of rainfall and temperature. The simplest of these is R. Lang's\textsuperscript{34} rain-factor, which is simply the ratio of the mean annual rainfall in millimeters to the mean annual temperature in degrees centigrade. This ratio can reach very high values when the mean annual temperature approaches zero and Lang proposed a modified ratio in which only the figures for those months having mean temperatures above zero are considered. This difficulty is avoided by E. De Martonne\textsuperscript{35} in his indice d'aridité, which is $P/(T+10)$, $P$ being rainfall in millimetres and $T$ mean temperature in degrees centigrade.

Another measure of the relative humidity of a climate is A. Meyer's NS quotient, which is the mean annual rainfall in millimetres divided by the mean deficit from saturation with water vapour expressed in millimetres of mercury. C. W. Thornthwaite\textsuperscript{36} has devised a quotient termed "precipitation effectiveness" given by the sum of the quotient for each month, which is defined as $115 \left(\frac{P}{T-10}\right)^{0.5}$ where $P$ is the mean rainfall in inches and $T$ the temperature in degrees Fahrenheit.

E. M. Crowther\textsuperscript{37} from a consideration of the effect of temperature and rainfall on leaching as shown by the silica-alumina ratio of the clay fraction found that to maintain constant intensity of leaching, a rise of $1^\circ$C. in temperature must be accompanied by an increase of 3.3 cm. in rainfall. He thus derives a measure of leaching termed the leaching factor, which is $R - 3.3T$, where $R$ is the mean annual rain-
fall in centimetres and $T$ is the mean annual temperature in degrees centigrade. This means that with a mean annual rainfall of 75 cm., or a little less than 30 inches and a mean annual temperature of about $23^\circ$C., the leaching factor would be zero. Leached soils could thus be defined as soils with a positive leaching factor and incompletely leached soils as soils with a negative leaching factor.

Many examples of correlations between soil properties are given by H. Jenny$^{18}$ who has made a careful study of North American data. In each case the correlation has been thrown into the form of an equation.

In studying the effect of humidity on soil properties, the most useful climatic constant would be one in which, as in Lang’s rain-factor or Meyers’ NS quotient the effect of temperature is taken into account. An example of this is the correlation between nitrogen content and NS quotient in the region east of the Rocky Mountains. Here, with a mean annual temperature varying around $11^\circ$C. the relationship can be expressed as

$$N = 0.320 (1 - e^{-0.0034Q}),$$

where $N$ is the nitrogen content per cent, $e$ the base of natural logarithms, and $Q$ the NS quotient. When the nitrogen content is correlated directly with rainfall the best equation is

$$N = 0.00655R - 0.023,$$

where $R$ is the mean annual rainfall in inches. Jenny also gives examples of correlation of other soil data with rainfall. A striking example of the effect of rainfall on a soil property is the correlation between rainfall and the depth of the calcium carbonate horizon in pedocalic soils. J. C. Russell and E. G. Engle$^{39}$ found the following figures for Nebraskan loess soils developed under varying rainfalls at a mean annual temperature of $48-50^\circ$F.

<table>
<thead>
<tr>
<th>Mean annual rainfall.</th>
<th>Depth of carbonate horizon.</th>
</tr>
</thead>
<tbody>
<tr>
<td>20 inches</td>
<td>12-24 inches</td>
</tr>
<tr>
<td>25 inches</td>
<td>35-47 inches</td>
</tr>
<tr>
<td>30 inches</td>
<td>about 60 inches</td>
</tr>
</tbody>
</table>
Temperature affects soil formation firstly by modifying the effectiveness of the rainfall. As we have seen, E. M. Crowther calculates that for each 1°C. rise in temperature an extra 3.3 cm. of rain is required to maintain constant percolation. Secondly, temperature affects all chemical processes in the soil and also all biological processes including both the growth of higher plants and the activity of microorganisms within the soil.

E. Ramann from a consideration of the relative degree of dissociation of water at different temperatures and an estimate of the number of days with temperatures above 0°C. in different regions worked out a weathering factor. For arctic regions he assumes that there are annually 100 frost-free days with a mean temperature of 10°C. At this temperature the relative dissociation of water, compared with unity at 0°C. is 1.7 and the weathering factor becomes 170. Similarly for mid-European conditions 200 days with a mean temperature of 18°C. and a relative dissociation of 2.4 gives 480 as the weathering factor. In tropical regions, 360 days at 34°C. and a weathering factor of 4.5 give a weathering factor of 1,620.

Whilst nitrogen and organic matter increase with rainfall, the effect of rising temperature is opposite, partly through its effect in diminishing the efficiency of the rainfall and partly through its effect in increasing the rate of decomposition. For regions of similar vegetation and rainfall Jenny gives an equation of the type

\[ N = Ce^{-kt} \]

where \( N \) = nitrogen per cent, \( T \) = temperature, and \( C \) and \( k \) are constants.

Whilst organic matter and nitrogen fall with increasing temperature, clay content increases, reflecting the increase in chemical weathering. With this increase in clay content there is the tendency for the molecular Si/Al and Si/R ratios to decrease to that under tropical conditions the typical clay will be of the ferrallitic type.
If we are to adhere to the distinction made between parent rock and parent material, i.e., weathered rock, we must admit that even although the rôle of the latter in the pedogenic processes is purely passive, yet the active factors of climate are involved by reason of the part played by them, acting through time, in determining the character of the weathered product. The ultimate passive ingredient is the parent rock. The parent rock through the weathered parent material can affect the final product in two ways, namely (1) through its physical properties and (2) through its chemical properties.

The most obvious way in which the parent rock can affect the resultant soil is through texture. Generally speaking acid igneous rocks, quartzose grits, and sandstones give rise to light sandy soils, which, except in depressed sites, tend to suffer from drought. On the other hand basic rocks and fine-grained sedimentary rocks tend to give heavier soils. Limestones give different types of soil according to their constitution and composition. Hard pure limestones tend to give light shallow soils, whilst impure soft limestones give deeper and heavier soils. Owing to the porous and jointed character of limestones, the resultant soils are generally dry.

The effect of parent rock on the chemical properties of the soil is chiefly through base-status. Acid rocks tend to give soils with low base-status and basic rocks soils with high base-status. The base-status affects the character of the weathering for, broadly speaking, high base status is associated with the formation of siliceous clay of the montmorillonite-beidellite type, whilst low base-status is generally associated with the formation of a kaolinitic product.

Base-status can also determine the type of eluviation. In a cool humid climate low base-status is associated with podsolization. On the other hand, with parent rock and parent material of high base-status podsolization does not occur. An example of this effect of parent material can be
seen on the one hand in the well developed podsols on acid quartzose sands in Southern England and on the other hand in the non-podsolized soils developed over basic igneous rocks in the cold humid climate of Iceland.

The effect of parent material is most plainly seen in the early stages of soil formation. Indeed, in regions of immature soils classification according to parent material may be entirely valid. Although with advancing development the effect of parent material is less marked, certain characteristics inevitably persist. Thus, in regions with such well-defined soils as podsols and chernozems the effect of parent material may be seen in texture, stoniness, and even possibly in colour. The effect of parent material on soil colour may be well seen in the soils of the Trias, Old Red Sandstone, and Red Measures (Coal Measures) in England. These soils are generally of a reddish colour. Yet, the colour is not the result of weathering or profile development but a rock character persisting in the resultant soils. The effect of parent material on soil formation is most markedly seen in the case of limestone. In many parts of the world, notably in the Mediterranean region, hard limestones give rise to a characteristic soil known as terra rossa. It is possible that even in Britain certain red soils over limestone may be related to this group. Contrasted with the red soils over limestone is a class of soils known as rendzinas. These are derived from soft limestones or calcareous clays and are characterized by the presence of free calcium carbonate throughout the profile, a strongly developed aggregate structure, and a dark type of humus. The mineral soil colour is generally greyish. These again are found in many parts of the world and are represented in Britain by certain dark-coloured soils of the chalk downs.

TOPOGRAPHY AS A PEDOGENIC FACTOR

Topography is a pedogenic factor mainly as affecting the operation of the climatic factors. Yet, it may be considered a direct factor in profile development where slope
determines the degree of erosion and may result in truncation.

Dealing first with the effect of topography on the climatic factors we note that it affects both temperature and the moisture-air regime in the soil. It affects temperature through elevation, aspect, and surface relief. The effect of elevation is fairly simple; there is approximately a fall of 1°F. for every 300 feet increase in altitude. This is reflected in a lower mean temperature, a shorter growing season and, because of decreased evaporation, a greater leaching efficiency of the rainfall. Thus, as compared with adjacent lowlands, upland regions have a colder and wetter climate.

The effect of aspect on temperature is easily understood. South slopes naturally receive more solar radiation and are, therefore, to that extent warmer, an effect that is often reflected in the type of vegetation and in the type of profile development.

Finally, there is the effect of relief. The most obvious effect of this is in the occurrence of so-called "frost-pockets" in basin-shaped areas. Although these effects may be of considerable importance for agriculture and horticulture, it is doubtful whether, as such, the occasional low temperatures occurring in clear still weather can have much effect on profile development.

The effect of topography on soil temperature and hence on soil profile is strikingly shown by some data obtained by H. Pallmann and E. Frei in a Swiss forest. Using a special technique for recording equivalent temperatures they obtained inter alia the following data:

<table>
<thead>
<tr>
<th>Height in metres</th>
<th>Slope</th>
<th>Aspect</th>
<th>Equivalent temperature at 10 cm. depth</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Winter</td>
</tr>
<tr>
<td>1910</td>
<td>30°</td>
<td>S.</td>
<td>3.2°C</td>
</tr>
<tr>
<td>1900</td>
<td>32°</td>
<td>NNE.</td>
<td>-0.8°C</td>
</tr>
</tbody>
</table>
The difference in temperature between two neighbouring localities at approximately the same altitude but differing only in aspect is very striking and is plainly reflected in the vegetation and in the soil profile. On the south slope is a calcareous soil, alkaline in reaction and low in humus, whilst on the north slope is an acid soil, carrying raw humus, and showing marked podsolization.

The effect of topography on the moisture-air regime in the soil is two-fold, namely, as affecting the actual rainfall and also as affecting its disposal. The effect on rainfall is very marked, the general tendency being for rainfall to increase with altitude as shown by the relief map. In Wales, the highest mean annual rainfall of about 200 inches is found near the peak of Snowdon (3,570 feet), whilst the lowest rainfall of about 25 inches are in open coastal lowlands. Similarly a rainfall map of England and Wales as a whole shows high rainfalls in the mountains and hills of the west and low rainfalls, occasionally down to less than 20 inches, in the flat coastal areas of the east. Apart from the altitude effect, there is, of course, a tendency for rainfall to decrease from west to east, the general direction of the rain-bearing winds.

Topography affects disposal of rainfall firstly by its effect on drainage conditions within the soil. Increase in altitude in itself tends to diminish evaporation since it is associated with cooler and wetter conditions. In microrelief, however, the effect is probably different for sites with elevated convex relief will tend to have greater evaporation than sites with a depressed concave relief.

The principal effect of topography, however, is on drainage conditions. In passing from a valley bottom to the adjoining upland the water-table tends to recede from the surface so that at the divide between one valley and the next it may be many hundreds of feet down and have little effect on water conditions in the actual soil profile. At or near the valley bottom the water-table is near the surface and dominates the water-air régime. We may thus have the apparent paradox that whilst the air climate is wetter in the upland
the soil climate is wetter in the lowland. This is shown also in the micro-relief, particularly in drift-covered country where dry soils stand up as knolls or low hills in an area of generally wet soils.

The effect of topography on erosion is most marked in regions such as the eastern United States where the climate and the character of the soil are such as to favour this process. In such regions slope is an important factor in determining both the character of the soil profile and its behaviour under cultivation. But even in a country such as Britain, where erosion is rarely catastrophic, surface relief has an important effect on soil depth, which is greater in lowlands and areas of flat or concave relief than in uplands and local areas of convex relief. The contrast is between the thin and often eroded soils of the uplands and convex sites and the deep soils of the lowlands and concave sites. Apart from the effect of relief on depth of profile there is also an effect on the leaching processes which are more intense with the shallower profiles. It is also possible that on slopes some of the leaching may be down the slope instead of vertical.

NATURAL VEGETATION AS A PEDOGENIC FACTOR

The connexion between vegetation and soil is so intimate that the principal soil groups can be almost completely defined according to the types of vegetation under which they occur. Broadly speaking, the character of the natural vegetation expresses the summation of the climatic factors under which it grows. Yet there are plant associations whose occurrence is determined mainly by soil conditions, whilst in other cases, the operation of climatic factors may be modified by the character of the soil. For example, forest is generally found under humid conditions. On the one hand, extreme humidity leads to the replacement of forest by bog or moor, whilst, on the other hand, with decrease of humidity, transition to steppe or savannah occurs. But, even in a region of typical forest
climate, local factors of topography and drainage may lead to the development of bog or moor vegetation. Similarly, in a steppe region, local conditions may lead to the development of forest.

The effect of soil conditions is seen most markedly in the transition zone between vegetation types, and may cause markedly different plant associations to occur under identical climates. An example of this is the former occurrence of steppe vegetation, now represented by chernozems, within the forest region of Western Europe.

And thus, whilst it is generally permissible to regard natural vegetation among the factors in soil formation, the reciprocal character of the contract between soil and plant should not be overlooked. Nor can vegetation be regarded as an independent factor, seeing that it is itself controlled by climate.* The operation of climate as a pedogenic factor is both direct on the soil itself and indirect through the vegetation.

The effect of forest as a vegetative cover is exerted mainly through the organic matter of the soil on which it occurs, and depends on the nature of the leaf-fall and the presence or absence of ground vegetation. In northern coniferous forests there is comparatively sparse ground vegetation, and the main source of soil organic matter is the leaf-fall, which, being naturally poor in mineral substances, tends to form a superficial layer of raw humus, favourable to the podsolizing processes. There are, however, differences among the conifers with respect to the mineral status of their leaf-fall.

*Locally, the apparently casual distribution of trees in a forest may affect profile development and appear to justify regarding vegetation as an independent factor. A. C. S. Wright (private communication) gives a striking example of this in the sub-tropical rain forest of New Zealand. Here, the kauri pine, Agathis australis, can cause local podsolization. The kauri is a long-lived tree (upwards of 800 years), and round the base of each tree there is built up a mound of bark fragments that form a thick mat of raw humus. Below this mat is a strongly leached lens-like A₄ horizon. Each tree has thus its own particular profile and the area forms a mosaic of these lens-shaped podsols.
Under deciduous forest, conditions are markedly different. Not only is the leaf-fall richer in mineral constituents, but there is generally a considerable contribution of organic matter from ground vegetation. The relatively high base-status of the litter and the residues from ground vegetation contribute to the formation of a neutral type of humus, whilst the presence of earth-worms ensures its distribution throughout the profile.

The organic matter profile of coniferous forest soils thus differs from that of deciduous forest soils. In the one case there is a sharply defined raw humus layer with a strongly acid reaction overlying a mineral soil that is relatively poor in organic matter. In the other case there is a moderate content of neutral or only slightly acid organic matter which decreases gradually in amount from the surface.

Premising that the excessive development of raw humus is disadvantageous to tree growth, even in coniferous forests, the importance of a clear view of the part played by soil, climate, and vegetation, is apparent. A soil developed from parent materials such as quartzose sand, or acid igneous rocks, can afford little compensation for the low mineral status of the leaf-fall of trees such as *Pinus sylvestris* and *Picea excelsa*, particularly if, owing to high rainfall and low mean temperatures, leaching is intense. G. Krauss, W. Worst, and G. Gärtner have exposed the deleterious effect of spruce plantings on a granite soil naturally poor in bases. Under such conditions, the best prospect of checking undue development of raw humus would be through the growth of trees whose leaf-fall has naturally a higher mineral status and which would therefore tend to conserve in the surface layer such base supplies as the soil possesses. It is possible that the encouragement of a ground vegetation by relatively open planting might also help to counter the tendency to raw humus development.

Under grassland or steppe conditions, the organic matter status of the soil is appreciably higher than under forest. This leads to the apparently paradoxical result that
in passing from a forest climate to the drier climate of steppe or prairie there is a marked increase in organic matter, which is maintained until the maximum content is reached in the middle of the chernozem belt. Thereafter, increasing aridity, by its effect in curtailing plant growth, is reflected in a decrease in organic matter.

The higher organic matter content of steppe soils compared with forest soils is due to a number of circumstances. Firstly, under forest, accessions of organic matter are mainly in the form of leaf-fall, which, accumulating on the surface, is exposed to serious losses by oxidative decomposition. Under steppe conditions, plant residues are added both at the surface, from the remains of leaves and stems of plants, and also within the soil, from dead plant roots. Secondly, the activity of earth-worms and other burrowing animals is greater under steppe than under forest. This ensures a more rapid and thorough incorporation of residues throughout the profile. Thirdly, summer drought under steppe conditions results in a restriction of biological decomposition to an extent not realized under the more humid forest conditions.

MAN AS A PEDOGENIC FACTOR

Over many parts of the world, and particularly in old settled countries like Britain, human interference has profoundly modified soils. This interference has operated mainly through the conversion of soils under natural vegetation, such as forest, prairie, or steppe, into agricultural land. There are also areas where the conversion has been from forest to pasture. In other areas the change has been from natural to artificial forest or from forest to scrub. Finally, there are areas where semi-desert or desert have been brought into cultivation by means of irrigation.

The effects of human interference in producing accelerated soil erosion have already been mentioned (page 93). Apart from these destructive effects, the conversion of soil from its virgin state to agricultural use results in many important modifications. The most obvious is the con-
founding of the original profile down to the depth of cultivation. Agricultural practices, including the application of liming materials and other fertilizers, may alter the general character of profile development. Podsolization may be repressed, and where cultivation has been deep enough to incorporate the upper part of the B horizon with the A horizon, the resultant profile may come to resemble a brown earth.

The effect of cultivation on organic matter status varies. Where deciduous forest has been replaced by mixed husbandry with the inclusion of grass leys the organic matter status is raised. On the other hand, permanent and exhaustive arable culture results in a decline in organic matter status and in certain conditions may make destructive erosion possible. The conversion of prairie or steppe to arable cultivation almost invariably implies a lowering of organic matter status and may be a precursor of disastrous wind erosion.

Whilst the initial effect of converting forest to arable is a lowering of organic matter status, the discontinuance of cultivation on account of erosion or soil exhaustion and the establishment of pasture for grazing may more than repair the losses in organic matter. Such a process has probably occurred in many of the hill lands of Britain, which passed through a stage of arable cultivation after the first clearance of the original forests.

Forestry practice may affect the course of profile development. Thus the replacement of deciduous by coniferous forest on soils with poor base status may intensify podsolization. The immediate effect of deforestation, by cutting down transpiration, may be to cause signs of drainage impedance to appear in the profile, whilst, on the other hand, a moderate degree of impedance may disappear on the establishment of forest. Finally, the selection of species in planting and their subsequent management may have important effects on the character of the forest humus, which in turn affects both the growth and profile development.
Fundamental changes in soil profile are made by such treatments as drainage, irrigation, and flooding.

Drainage may change a profile with impeded leaching to a leached profile. Where, however, the impedance is due to the character of the parent material some of the original characters will persist. Some degree of impedance is always present in clay soils, however well drained. The improvement consequent on drainage depends on its maintenance and, with neglect, reversion to the original state occurs, as may be seen in large areas in Britain formerly improved by tile drainage.

Irrigation can have different effects on the soil according to the character of the irrigation water and the way in which it is used. Generally speaking, soils used for irrigation are desert or semi-desert soils, low in organic matter, of high base-status, and in many cases with a relatively high salt content. Where irrigation is accompanied by drainage and where irrigation waters are low in sodium salts, the excess salinity is removed. On the other hand, where no drainage is provided, and particularly where the irrigation waters themselves contain sodium salts, the soil becomes progressively salinized and may eventually become unfit for the growth of crops. Large areas of land formerly carrying crops under irrigation and now derelict, may be found in many parts of the world. (See Chapter XIII.)

Flooding accompanied by deposition of silt from flood water builds up an artificial soil profile. Examples of this are the warp soils round the Humber estuary in England and the soils built up from Nile alluvium in Egypt. Equally artificial are the soils formed by endyking of sea marshes, such as the Dutch polder soils and the soils of Romney Marsh in Kent.

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CHAPTER V

THE CLAY COMPLEX

We have seen that soil may be conveniently considered as being composed of a relatively inert framework of unweathered minerals, together with the so-called clay-humus complex, consisting mainly of the products of the chemical weathering of silicates and with humus, which is the product of the decay of plant and animal residues in the soil. This association is probably not simply that of a mixture of components but quasi-chemical in character. Indeed, L. Meyer has actually prepared from artificial permutites and humic acids complex bodies similar to the clay-humus of chernozems. Mayer considers the organic component to be an integral part of the structure and not superficially absorbed. He did not succeed in preparing complexes from clay minerals but considers that they may arise in nature by the association in statu nascendi of humus formed from plant residues and clay minerals formed from the weathering of primary silicates.

Although in the soil the clay-humus complex may be considered to behave as a single entity, certain constituents can be separately identified. Further, the study of soil profiles involves the study of horizons in which the complex is almost entirely mineral in character. It is therefore convenient for the purposes of presentation to consider the mineral part of the clay complex as if it were an independent constituent. Such a study is necessary in order to understand the behaviour of the complex as a whole. In the same way, humus may be considered as a separate entity although it is to a large extent associated or combined with the clay.
RELATIONSHIP OF CLAY COMPLEX TO WEATHERING-COMPLEX AND CLAY FRACTION

In considering the mechanical composition of the soil, we had occasion to distinguish the so-called clay fraction, which was defined in terms of particle-size or, more correctly, settling-velocity. We may also recognize an inorganic colloidal fraction, or complex, which is the reactive mineral constituent of the soil and which confers on it those physical properties that differentiate soil from physically comminuted powders. Finally, we recognize a certain portion of the soil as consisting of material of secondary origin that has resulted from the chemical weathering of silicate minerals. Can we regard the three groups thus defined as being identical? In other words, can we assume that the clay fraction contains all the chemically reactive colloidal inorganic material of the soil, and that this is identical with the fraction which has resulted from the chemical weathering of mineral silicates?

Whilst we cannot assume the complete identity of these groups, there is probably no great error, for ordinary soils, in the assumption. Certain soils may, indeed, contain material of secondary origin in the fractions other than the clay. This is particularly the case in certain tropical soils where the coarse material may be almost entirely secondary. Coarse concretionary material may also occur in certain illuvial horizons and in horizons developed in proximity to a water-table. But such secondary material, though strictly belonging to the weathering-complex, would scarcely exhibit chemical reactivity or confer on the soil the characteristic physical properties of colloids. Apart from such cases as these, the material coarser than clay is generally composed of original crystalline unweathered minerals and rock fragments. We may assume, therefore, with the exceptions mentioned, that the clay fraction includes all the secondary products of weathering—the weathering-complex.

It is less easy to satisfy ourselves that the clay fraction as defined in mechanical analysis (<0.002 mm.) does not
contain unweathered minerals. Crystalline material is certainly present. The curious satin-like appearance sometimes observed in agitated suspensions of clay fraction is due to reflection from small platelets. X-ray analysis has, indeed, revealed the crystalline character of most of the clay fraction. This micro-crystalline material is, however, of secondary origin, and we are more concerned to know whether original unweathered minerals are present.

A. F. Joseph found that the whole of the clay fraction isolated from certain Sudan soils could be further dispersed into "ultra-clay" and was presumably of uniform character. S. B. Hendricks and W. H. Fry found no felspar or mica below 0.001 mm. (1μ). R. Bradfield found no essential difference between "suspensoid" clay (0.4-1.2μ) and "emulsoird" clay (<0.1μ) from a Missouri soil.

On the other hand, A. Atterberg concluded that biotite forms the plastic ingredient of certain Swedish clays, whilst hæmatite and limonite are present in the clay of terra rossa. He found that other lamellar minerals, such as kaolinite, talc, and muscovite, can exhibit the clay property of plasticity when sufficiently comminuted.

From a consideration of mechanical composition curves of physically comminuted materials, G. W. Robinson concluded the lower limit of physical subdivision to be represented by particles of approximately 0.6μ diameter (log V=5.0). Data for the composition of successive fractions of typical soils also suggest that, except in certain highly weathered tropical soils, unweathered minerals are present in the fractions above 0.6μ diameter.

A study by mineralogical and optical methods of the successive fractions of different soils by C. E. Marshall showed that primary minerals may occur in the coarser part of the fraction below 2μ (0.002 mm.) diameter. He would prefer 1μ as the upper limit of diameter for the clay. Data obtained by I. C. Brown and H. G. Byers on the chemical composition of successive fractions down to 0.1μ appear to support this conclusion. The present writer found that

*Unpublished data.
whilst in soils of primary weathering, successive fractions of the clay were similar in composition, the material at the upper end of the clay range in soils of secondary weathering was much more siliceous than the finer material. Even for soils of primary weathering, an upper limit of log $V=4.0$ may be preferable to the limit log $V=4.5$ of the international scale. For most purposes the error with soils of primary weathering would not be great. With soils of secondary weathering, the upper limit should be lower to ensure the exclusion of unweathered minerals. It seems possible, therefore, that the clay fraction as ordinarily defined may contain some unweathered physically comminuted minerals in the upper part of its range, but the proportions thus present would generally be small. Although, in the writer's view, it would have been better to fix the clay limit lower than $2\mu$, there is probably no serious error, for most purposes, in identifying the clay fraction with the weathering-complex. The error is likely to be most serious in soils of secondary weathering and soils in which the weathering has been mainly physical. The presence of concretionary material would also introduce errors. To fix the limit higher would be to include unweathered minerals, whilst to fix it lower would exclude secondary material. In most cases there will be a certain overlap and a hard and fast separation is impossible.

EARLIER VIEWS ON THE NATURE OF CLAY

Premising that the clay complex is generally and mainly the product of the chemical weathering of silicate minerals, its constitution has been variously pictured.

The earliest theory, still extant in many textbooks, regards kaolinite as the essential mineral of clay, and chemical weathering as the process of kaolinization, represented ideally, in the case of orthoclase felspar, by the equation:

$$K_2O\cdot Al_2O_3\cdot 6SiO_2 + 3H_2O =$$
$$Al_2O_3\cdot 2SiO_2\cdot 2H_2O + 2KOH + 4SiO_2$$
It should be remarked, however, that the material kaolin, whose characteristic mineral is kaolinite, is the product of deep-seated hydrothermal decomposition, which differs entirely from the epigene processes whereby soils are formed, although these processes may, under certain conditions, yield minerals of the kaolinitic group. Further, it is deficient in the properties of plasticity, colloidal imbibition, and base exchange, which are characteristic of soils. As will be seen later, kaolinite is only one of a number of minerals found in clay and formed from primary minerals by chemical weathering. The kaolinization theory is thus inadequate as a general explanation of the origin of the clay complex, although the clay complex may contain kaolinite or minerals similar in constitution to kaolinite, formed by epigene processes.

J. M. Van Bemmelen⁹ considered the weathering-complex of the soil to consist of two fractions, namely, a complex-A, soluble in hot concentrated hydrochloric acid*, and having a molecular SiO₂/Al₂O₃ ratio varying from 3 to 6, and a complex-B, soluble only in hot concentrated sulphuric acid, having a molecular SiO₂/Al₂O₃ ratio varying from 2 to 3. The complex-A was considered to be an absorption compound of indefinite composition and colloidal character, whilst complex-B was considered to be essentially similar to kaolin and to be less reactive than complex-A. D. J. Hissink, J. Van der Spek, and S. B. Hooghoudt¹⁰ have recently used Van Bemmelen’s methods in the examination of certain Dutch soils. They found about 70% of the clay fraction to consist of complex-A, which was responsible for about 90% of the chemical reactivity as measured by base-exchange capacity (see p. 159). On the other hand R. H. J. Roborgh and N. H. Kolkmeyer¹¹ found that the extraction

*In methods depending on extraction with strong mineral acids, the liberated silicic acid is extracted by a subsequent treatment with warm dilute alkali. Different methods are prescribed for this alkaline extraction, but the results cannot be regarded as highly trustworthy since there is always the possibility of some original quartz being attacked.
of a clay soil with hydrochloric acid by Van Bemmelen's method had little effect on base exchange capacity and clay content.

H. Stremme\textsuperscript{12} distinguishes three groups of hydrous aluminium silicates formed by the weathering of mineral silicates. The first group are considered to be residual products from the decomposition of felspars and felspathoids, and approximate in composition to kaolinite, \( \text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 2\text{H}_2\text{O} \). Though not demonstrably crystalline, they do not form gels. The members of the second group, the allophanoids, have \( \text{SiO}_2 / \text{Al}_2\text{O}_3 \) ratios varying from less than 2.0 to greater than 2.0. By replacement of aluminium with iron, corresponding ferric silicates are formed such as nontronite, chloropal, etc. They are typical gels and are formed partly as precipitates and partly as residual products from the decomposition of ferromagnesian minerals. H. Harrassowitz\textsuperscript{13} considers them to originate also from lime felspars. The third group includes the zeolites, which do not apparently occur in soils, but represent the crystalline prototypes of the allophanoids.

R. Ganssen\textsuperscript{14}, from the results of acid extraction designed to separate the weathering-complex from the unweathered portion of the soil, postulates a compound of the formula, \( \text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2 \cdot R''\text{O}(R'\text{O}) \), and accounts for actual variations from this formula by the presence of excess of silica or excess of alumina. It may be doubted, however, if any method based on acid extraction can effect a separation of the clay complex from the unweathered portion of the soil, for certain minerals have an appreciable solubility even in dilute acids. For example, a powdered quartz-diorite, in which, of course, no weathered constituent was present, was found by the writer to lose 10.8\% silica, 20.6\% alumina and ferric oxide through extraction with 20\% hydrochloric acid at 100° C. Acid extraction may be applicable to soils consisting mainly of clay complex and quartzose material.

A further defect of the earlier theories as to the constitution of the clay or weathering-complex of soils is that they
were generally conceived in terms of compounds or complexes built up essentially of silica and alumina, with iron oxides as adventitious constituents. It is now known that iron compounds form essential ingredients in the clay complex, for clays containing notable proportions of iron occur in which the presence of adventitious ferric oxide cannot be assumed. Any theory of the nature of the clay complex must take account of the possible existence of complex ferro- or ferri-silicates and of the possibility of isomorphous replacement of aluminium by iron in complex aluminosilicates.

The similarity of the base exchange reactions of the soil to those exhibited by zeolites has led certain workers to postulate the presence of these minerals in the weathering complex. The clay complex has, in fact, been termed by some writers the zeolithic (or zeolitic) complex. But since none of the known zeolites has been proved to be present in soils, it appears better to avoid the use of this term.

COMPOSITION OF THE CLAY FRACTION

If it is admitted that the clay fraction can be assumed to represent the weathering-complex, the study of its composition may be expected to throw light on the character of the weathering processes. During recent years, this method of attack has been increasingly used and valuable results have been obtained. Some data for the chemical composition of the clay fraction of typical soils are given in Table VI. It should be observed that, whilst in most cases the fraction <0.002 mm. has been used, the data of W. O. Robinson and R. S. Holmes^15 relate to the clay colloid or ultra-clay (<0.3μ). From what has been said above they may be regarded as comparable. It will be seen that the composition of the clay shows a very wide variation, and that in all cases it departs from the composition of kaolinite, once supposed to be the characteristic ingredient of clay. Two explanations are possible. Either the clay fraction consists of a kaolin-like mineral with excess of sesquioxides, or in some cases, silica; or the minerals present form a mixture of hydrated aluminosilicates and
ferrisilicates of varying composition, mixed in some cases with excess of sesquioxides or silica.

**Table VI.—Composition of Clay Fractions of Soils**

<table>
<thead>
<tr>
<th>SOIL</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>Fe₂O₃</th>
<th>TiO₂</th>
<th>CaO</th>
<th>MgO</th>
<th>K₂O</th>
<th>Na₂O</th>
<th>Authority</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>44.89</td>
<td>22.59</td>
<td>7.75</td>
<td>0.47</td>
<td>1.48</td>
<td>1.44</td>
<td>1.36</td>
<td>0.22</td>
<td>W. O. Robinson and R. S. Holmes¹⁵</td>
</tr>
<tr>
<td>II</td>
<td>33.95</td>
<td>36.06</td>
<td>11.02</td>
<td>0.62</td>
<td>0.31</td>
<td>0.40</td>
<td>0.56</td>
<td>0.44</td>
<td>R. S. Holmes¹⁵</td>
</tr>
<tr>
<td>III</td>
<td>53.34</td>
<td>20.77</td>
<td>7.15</td>
<td>0.81</td>
<td>5.01</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n. d.</td>
</tr>
<tr>
<td>IV</td>
<td>49.5</td>
<td>30.9</td>
<td>12.2</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>A. D. Hall¹⁶</td>
</tr>
<tr>
<td>V</td>
<td>44.08</td>
<td>27.64</td>
<td>21.81</td>
<td>n.d.</td>
<td>0.58</td>
<td>1.61</td>
<td>1.10</td>
<td>0.96</td>
<td>J. Hendrick and W. G. Ogg¹⁷</td>
</tr>
<tr>
<td>VI</td>
<td>43.9</td>
<td>35.6</td>
<td>19.9</td>
<td>n.d.</td>
<td>0.3³</td>
<td>0.4</td>
<td>0.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>VII</td>
<td>58.9</td>
<td>17.3</td>
<td>20.8</td>
<td>n.d.</td>
<td>0.2</td>
<td>2.4</td>
<td>1.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>VIII</td>
<td>46.6</td>
<td>33.3</td>
<td>12.6</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>B. de C. Marchand and C. R. Van der Merwe¹⁸</td>
</tr>
<tr>
<td>IX</td>
<td>10.19</td>
<td>15.84</td>
<td>62.51</td>
<td>0.20</td>
<td>0.23</td>
<td>0.05</td>
<td>tr.</td>
<td>G. W. Robinson¹⁹</td>
<td></td>
</tr>
<tr>
<td>X</td>
<td>27.6</td>
<td>37.8</td>
<td>26.8</td>
<td>1.4</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>G. W. Robinson¹⁹</td>
<td></td>
</tr>
</tbody>
</table>

I. Carrington Loam, 0-9", Iowa, U.S.A.
II. Cecil Clay, 0-9", Georgia, U.S.A.
III. Houston Black Clay, 0-12", Texas, U.S.A.
IV. Average of 8 soils from Rothamsted, England. Ignited.
V. Craibstone, Scotland. Ignited.
VI. Transvaal Red Loam. Ignited.
VII. Transvaal Black Turf. Ignited.
VIII. Shale Soil, N. Wales. Ignited.
IX. Nipe Lateritic Clay, Cuba.
X. Red Loam, Tanganyika. Ignited.

Robinson and Holmes (loc. cit.), reviewing the results of their analyses of the "clay colloid" from a range of American soils, examined the hypothesis that the primary product of chemical weathering was a mixture of a hydrated aluminium silicate, \(\text{Al}_2\text{O}_3.2\text{SiO}_2.x\text{H}_2\text{O}\) of the kaolinite type, and a hydrated ferric silicate, \(\text{Fe}_2\text{O}_3.2\text{SiO}_2.x\text{H}_2\text{O}\). They found it necessary in the majority of cases to assume the presence of an excess either of silica or sesquioxides.

G. W. Robinson²¹, from a study of the composition of the clay fraction of soils in Wales, concluded that, for soils of primary weathering, the data were not incompatible with the theory that the first product of weathering under humid temperate is a mixture of hydrated silicates of the general formula, \(\text{R}_2\text{O}_3.2\text{SiO}_2.x\text{H}_2\text{O}\). Variations from the \(\text{SiO}_2/\text{R}_2\text{O}_3\) ratio of 2.0 demanded by this hypothesis could be ascribed to
differentiation by eluviation or, in the case of soils derived from unconsolidated sediments, concomitant precipitation of silicic acid from river waters. Thus; whilst the average molecular $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio for the clay fraction of soils directly formed by primary weathering was 1.85, the corresponding ratio for the clay fraction of soils derived from unconsolidated aqueous sediments was 2.67. Analyses were also made of the clay fractions from twelve samples of material of primary weathering (C horizons) in N. Wales, lying below the ordinary soil horizons. The following results were obtained:

<table>
<thead>
<tr>
<th></th>
<th>Range</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mol. $\text{SiO}_2/\text{Al}_2\text{O}_3$</td>
<td>2.28—3.54</td>
<td>2.61</td>
</tr>
<tr>
<td>Mol. $\text{SiO}_2/\text{R}_2\text{O}_3$</td>
<td>1.96—2.16</td>
<td>2.03</td>
</tr>
</tbody>
</table>

The close approximation of the average composition to a figure represented by a molecular $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio of 2.0 seems significant.

It should be noted that, although silica and sesquioxides account for most of the clay fraction, there are also present alkali and alkaline earth bases. These are present partly as reactive or exchangeable cations and partly as non-exchangeable cations. Their significance will be appreciated at a later point in this account.

**CRYSTALLINE CHARACTER OF CLAY COMPLEX**

The earlier writers on soils, in discussing the nature of the clay constituent, generally assumed that it has the character of an amorphous gel, but later work brought into question this view of the character of clay. For example, E. B. Powell\(^{22}\) questioned the emulsoid character of colloidal clay on the ground of the smoothness of the curves connecting the viscosity of clay suspensions with temperature.

Further work has shown that, as in the case of many other gels, the constituent material is principally micro-crystalline. The X-ray studies of C. S. Ross\(^{23}\) revealed the crystalline structure of certain clays and also permitted the identification of some definite minerals as common constituents of clays. The clays examined by Ross were divided by him into two
groups, namely, (1) the kaolin group, including kaolinite, \(\text{Al}_2\text{O}_3\cdot2\text{SiO}_2\cdot2\text{H}_2\text{O}\) and anauxite, \(\text{Al}_2\text{O}_3\cdot3\text{SiO}_2\cdot2\text{H}_2\text{O}\); and (2) the montmorillonite group, including montmorillonite, \((\text{Ca},\text{Mg} \cdot \text{O}).\text{Al}_2\text{O}_3\cdot5\text{SiO}_2\cdot5\text{H}_2\text{O}\), beidellite, \(\text{Al}_2\text{O}_3\cdot3\text{SiO}_2\cdot0.4\text{H}_2\text{O}\), nontronite, \(\text{Fe}_2\text{O}_3\cdot3\text{SiO}_2\cdot0.4\text{H}_2\text{O}\), and crystalline halloysite, \(\text{Al}_2\text{O}_3\cdot2\text{SiO}_2\cdot0.3\text{H}_2\text{O}\). The dominant soil- and clay-forming mineral was considered by Ross to be beidellite, together with its ferric isomorph nontronite. These minerals exhibit a platy crystalline form analogous to mica, and it may be maintained that this structure, involving large surface development and relatively large contact areas between particles, is sufficient to explain the characteristic physical properties of clays. The crystalline nature of clay particles is not inconsistent with their being built up into a gel structure having the property of imbibing water. The structural elements of many other gels are also known to be crystalline.

S. B. Hendricks and W. H. Fry\(^{24}\) applied the methods of X-ray examination to the colloidal material isolated from a wide range of soils and clays. The crystalline character of the materials examined was shown by comparison with the diffraction patterns obtained from known clay minerals, and it was possible to characterize each of the samples examined. In some cases, notable quantities of quartz were shown to be present, whilst a few samples gave kaolinite patterns. In many cases, montmorillonite, \(\text{Mg} (\text{Ca}) \cdot \text{O} \cdot \text{Al}_2\text{O}_3\cdot5\text{SiO}_2\cdot n\text{H}_2\text{O}\) or beidellite, \(\text{Al}_2\text{O}_3\cdot3\text{SiO}_2\cdot n\text{H}_2\text{O}\) were indicated; but since these minerals give identical diffraction patterns, it was not possible to distinguish them. Bentonite, a mineral similar to montmorillonite, was present in certain samples, and halloysite, a mineral similar in composition to kaolinite, in others. Hendricks and Fry distinguished three main types of clay, namely, (1) the montmorillonite type; (2) the halloysite type; and (3) the Ordovician bentonite type. Colloids with high iron content frequently gave the halloysite pattern. It was presumed that in each type some isomorphous replacement of aluminium by iron occurred. Bauxite was found in one sample, whilst it was
impossible to identify the minerals of a lateritic clay from Cuba consisting mainly of sesquioxides.

One point which emerged from the work of Hendricks and Fry was that the primary soil minerals, such as felspars and micas, were not found among the crystalline material of the colloidal fraction, a fact that suggests that the lower limit of their physical sub-division lies above the limit used (0.001 mm.) in defining the colloidal material for this work.

C. E. Marshall\textsuperscript{25} adduced evidence for the crystalline character of clays from a study of their double-refraction properties, and showed that the univalent and bivalent cations associated with clays are not, as was supposed by some workers, present at the surface alone, but can occupy definite positions in the crystal lattice. From ultra-microscopical observations, he has concluded that, for a given clay, each cationic combination has a characteristic degree of dispersion.

W. P. Kelley, W. H. Dore, and S. M. Brown\textsuperscript{26} compared potash bentonites with colloidal material from certain soils and found similar X-ray patterns.

C. E. Marshall\textsuperscript{27}, in an examination of sub-fractions of three clays by optical methods, found the following minerals present:

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Bentonite</th>
<th>Putnam Clay</th>
<th>Rothamsted Clay</th>
</tr>
</thead>
<tbody>
<tr>
<td>$2\mu-1\mu$</td>
<td>$\beta$-Cristobalite and montmorillonite</td>
<td>Quartz and bréidellite</td>
<td>Quartz and bréidellite</td>
</tr>
<tr>
<td>$1\mu-0.5\mu$</td>
<td>ditto</td>
<td>ditto</td>
<td>Fe-rich bréidellite and halloysite</td>
</tr>
<tr>
<td>$0.5\mu-0.2\mu$</td>
<td>Montmorillonite</td>
<td>Beidellite</td>
<td>ditto</td>
</tr>
<tr>
<td>$0.2\mu-0.1\mu$</td>
<td>ditto</td>
<td>ditto</td>
<td>Fe-rich bréidellite</td>
</tr>
<tr>
<td>$&lt;0.1\mu$</td>
<td>ditto</td>
<td>ditto</td>
<td>ditto</td>
</tr>
</tbody>
</table>

During recent years a considerable amount of work has been carried out on the actual lattice structure of clay minerals. According to the ideas of L. Pauling\textsuperscript{28} and others, based on the theories of W. L. Bragg, it is supposed that the clay minerals have crystal lattices built up of successive layers of linked atoms. Two types of layer occur, namely (1) the silica layer consisting of an hexagonal network of
tetrahedra, each formed by a central silicon atom with three oxygen atoms shared with adjacent silicons and one free oxygen. The composition is represented by \([\text{Si}_2\text{O}_5^-]_n\) the two negative charges being due to the unshared oxygen and available for the attachment of other layers; and (2) the hydrargillite or gibbsite layer \((\text{Al}_2(\text{OH})_6)n\), each alumi-
nium atom (co-ordination number 6) being at the centre of an octahedral arrangement of hydroxyl groups; each octahedron shares three edges with adjacent octahedra. Magnesium can also occur in octahedral arrangement.

The unit structures of the clay minerals are obtained, in the simplest case, by condensation of silicon tetrahedra with hydrargillite octahedra whereby each vertical oxygen atom of the silica layer replaces two shared hydroxyls of the hydrargillite layer. We thus have

$$\text{Al}_2(\text{OH})_4^{++} + \text{Si}_2\text{O}_5^{-} = \text{Al}_2\text{Si}_2\text{O}_5(\text{OH})_4$$

(kaolinite)

and where a similar condensation takes place on the other side of the hydrargillite layer

$$\text{Al}_2(\text{OH})_2^{+++} + 2\text{Si}_2\text{O}_5^{-} = \text{Al}_2\text{Si}_4\text{O}_{10}(\text{OH})_2$$

(pyrophyllite) (Fig. 5).

The former type of structure may be briefly referred to as the Si-Al type and the latter as the Si-Al-Si type.

A number of substitutions are possible. In the first place Al$^{+++}$ may be replaced by Fe$^{+++}$ or Fe$^{++}$. In the latter case there is an unsatisfied valency, which may be satisfied by a univalent ion. The second important substitution is that whereby silicon is replaced by aluminium in 4 co-ordination, i.e. in tetrahedral form. Here again the result of such replacement is that there is an unsatisfied valency, which can be compensated by univalent or bivalent ions. Thirdly, octahedral aluminium may be replaced by octahedral Mg., the resultant charge being balanced by a cation equivalent, or 2Al is replaced by 3Mg. In addition to these replacements, C. E. Marshall$^{28}$ suggests the possibility of replacement of Al by Ti and of Si by P, each involving changes in charge.

The Si-Al or Si-Al-Si layers are ideally infinite in extent and each unit stack or cell is connected with the rest by weak secondary valencies, a circumstance usually associated with platy or flaky cleavage. On the other hand the layers in each stack are held together by strong ionic bonds. The successive unit layers may show randomness in stacking;
for this reason large well-developed crystals are not formed in this group of minerals.

The clay minerals occurring in soils fall into a number of types.

(a) **Kaolin group.** These minerals are of the Si-Al type and correspond with the general formula $\text{Al}_2\text{O}_3.\text{SiO}_2.\text{nH}_2\text{O}$. Within the group are distinguished (i) the kaolinite group, corresponding with $\text{Al}_2\text{O}_3.2\text{SiO}_2.2\text{H}_2\text{O}$; (ii) the anauxite group, similar to the kaolinites but with Si/Al ratio greater than 2.0; and (iii) the hydrated halloysite group, corresponding with the formula $\text{Al}_2\text{O}_3.2\text{SiO}_2.4\text{H}_2\text{O}$. By far the most important of these from the standpoint of soil formation is kaolinite. The basal spacing, i.e., the thickness of the unit cell is about 7Å ($\lambda=10^{-8}$ cm.) for kaolinite and 10Å for hydrated halloysite. The Si/Al ratio varies from 1.8 to 2.1 in different members, owing to limited isomorphous replacement of $\text{Al}^{+++}$ by $\text{Fe}^{+++}$ and $\text{Si}^{+++}$ by $\text{Al}^{+++}$. Viewed under the electron microscope kaolinite appears as hexagonal plates, whilst halloysite shows rod-like forms.

The kaolin minerals show low base exchange capacity, but this may be increased by comminution, i.e., by increasing the proportion of broken edges.

(b) **The pyrophyllite-montmorillonite group.** The minerals of this group are of the Si-Al-Si type and are all derived from the ideal mineral pyrophyllite $\text{Al}_2\text{O}_3.4\text{SiO}_2.\text{H}_2\text{O}$ or $(\text{Al}_2)(\text{Si}_4)\text{O}_{10}(\text{OH})_2$. In the actual minerals (i) octahedral aluminium may be replaced by $\text{Fe}^{+++}$ or by octahedral Mg, in the former case without change of valency but in the latter case with the development of a negative charge; and (ii) silicon may be replaced by tetrahedral aluminium with the development of a negative charge. The general formula may be given as:

$$(\text{R}_2^{+++}\text{R}_n^{+++})\text{(Al}_m^{+++}\text{Si}_4^{+++})(\text{OH})_{10}\text{xH}_2\text{O},$$

xH$_2$O representing interplanar water. A typical montmoril-
Ionite would be given by the substitution of one-sixth of the Al by Mg and the satisfaction of the resulting valency by an equivalent amount of Na giving

\[(\text{Al}_{1.67}\text{Mg}_{0.33})(\text{Si}_4\text{O}_{10}(\text{OH})_2)]\text{Na}_{0.33}\.

By replacement of silicon by tetrahedral Al, beidellite is obtained. According to C. S. Ross and S. B. Hendricks\(^{30}\) this may be represented as \((\text{Al}_{2.17})(\text{Al}_{0.83}\text{Si}_{3.17})\text{O}_{10}(\text{OH})_2\).

\[
\downarrow
\]
\[\text{Na}_{0.33}\]

Many minerals intermediate between montmorillonite and beidellite occur.

Ferric iron can replace octahedral aluminium in this series, giving nontronites. An actual example given by Ross and Hendricks (from Sandy Ridge, N. Carolina) can be expressed by the formula:

\[(\text{Al}_{0.39}\text{Fe}^{+++}_{1.70})(\text{Al}_{0.58}\text{Si}_{3.42})\text{O}_{10}(\text{OH})_2\]

\[
\downarrow
\]
\[\text{H}_{0.31}\]

There are many examples of clays of the beidellite-montmorillonite group with smaller amounts of iron than would justify the description of nontronite.

N. Hoffmann, K. Endell, and W. Wilms\(^{31}\) showed that beidellite and montmorillonite give the same X-ray diffraction pattern, that these minerals have a wide spacing between the layers, and that the spacing varies with the moisture content. S. B. Hendricks, N. A. Nelson and L. T. Alexander\(^{32}\) have studied the water content and interlayer spacing of montmorillonites. In calcium montmorillonites up to three water layers may separate each unit, giving a spacing of 18Å; with two water layers the spacing is about 15Å. Samples dried at 200°C. give a spacing of about 16Å.

(c) The hydrous mica or illite group. Minerals of this group were described as Ordovician bentonite by Hendricks and Fry, as Tonmineral–X by A. Jacob, V. Hoffmann, H. Loofmann, and E. Maegdefrau\(^{33}\), and as illite by R. E.
Grim, R. H. Bray and W. F. Bradley. They resemble the pyrophyllite minerals in being built up on the Si-Al-Si pattern, by substitution of one of four silicons by tetrahedral aluminium and with introduction of potassums as compensating ions. The potassium ions lie in sheets between the double layer units. The basal spacing is about 10Å and does not, as in the case of montmorillonite, vary with the moisture content. Isomorphous substitutions among the octahedral ions are common.

(d) Palygorskite group. Minerals of this group are less common than those of the other groups. The best known member is attapulgite. According to W. F. Bradley it is based on double chains of silicon tetrahedra and the chemical constitution of a unit is \((\text{OH}_2)_4(\text{OH})_2\text{Mg}_2\text{Si}_8\text{O}_{20}\text{Si}_2\text{HO}_2\). It is essentially a chain-like arrangement of silica and brucite layers with associated water.

C. E. Marshall gives the following schematic classification of the clay minerals:

- **Silicate basis of structure**
  - SiO₅
    - Fibrous clays with Amphibole-like structure
      - (Attapulgite, Sepiolite)
    - Complete sheets
  - SiO₅
    - 2:1 lattice type
      - Transition group between micas and clays
        - (Illite—Bravaisite)
        - (Vermiculites)
        - (Micaceous halloysite)
    - Expanding lattice clays
      - Hydrated (Halloysite)
      - Dehydrated (metahalloysite)
        - (Kaolinite)
        - (Nacrite)
        - (Dickite)
    - Net charge is chiefly due to replacement in centre gibbsite sheet
      - (Montmorillonite-Saponite)
    - Net charge is chiefly due to replacement in silica sheets
      - (Beidellite-Nontronite)
Apart from the main types above-mentioned, mixed structures are possible. S. B. Hendricks and L. T. Alexander\textsuperscript{37} consider that structures may occur with alternating layers of montmorillonite and mica type. Analogous structures have been found in other mineral groups.

In addition to the mixed mineral structures mentioned it is very probable that there are present in soils structures in which organic groups enter into the lattice structure. Reference has already been made to the suggestion of L. Meyer\textsuperscript{38} that mixed organo-mineral structures can originate during the weathering of primary silicates in the presence of humifying organic matter.

**IDENTIFICATION OF CLAY MINERALS**

(a) \textit{X-ray diffraction}. This method depends on the determination of the diffraction patterns obtained when monochromatic X-rays are allowed to fall on samples of crystalline material. Each mineral gives a set of reflections which are characterized by their spacing, width, and relative intensities. For details of the methods the reader may be referred to standard works on crystallography. The X-ray diffraction method can give an indication of the minerals and an approximate estimate of the relative proportions present.

(b) \textit{Dehydration curves}. These methods depend on the loss in weight of samples heated to different temperatures. The loss in weight is plotted against the temperature and characteristic curves are obtained. An example of the type of curves obtained by W. P. Kelley, H. Jenny, and S. M. Brown\textsuperscript{39} is given in Fig. 6.

(c) \textit{Thermal analysis}. This method, originally due to Le Chatelier, and applied to soil material by V. Agafonoff and G. Jouravsky\textsuperscript{40}, is based on the recording of endothermic and exothermic changes during the heating of clay minerals. A sample of the mineral is heated up in a furnace side by side with a similar sample of an inert material. The temperature of the furnace is recorded as well as the difference in temperature between the sample and the inert
material. When an exothermic change takes place the sample shows a higher temperature than the inert material, whilst the reverse is the case for an endothermic change. Kaolinite shows a marked endothermic change at 550-600°C. and an exothermic change at 950-1,000°C., corresponding

![Graph: Dehydration curves of montmorillonite and kaolinite.]

Fig. 6.—Dehydration curves of montmorillonite and kaolinite.

with loss of water and crystal rearrangement respectively. Montmorillonites and illites show endothermic losses of water at about 200°C. and at about 700°C. By measuring the size of the peaks and troughs on the heating curve a quantitative estimate of the amount of a given mineral may be estimated. The method has been adapted for the detec-

(d) Dye absorption. Dye-absorption tests for distinguishing clay minerals have been proposed from time to time, notably by J. B. Page, who found the methods somewhat untrustworthy.

(e) Electron microscope. By this instrument it has been found possible to identify clay minerals. Kaolinite shows hexagonal plates, and halloysite rod-shaped forms. Montmorillonite and illite are platy, whilst nontronite is rod-shaped. The method has been used successfully by C. E. Marshall, R. P. Humbert, and B. T. Shaw.

(f) Base exchange properties. The capacity of a mineral to exchange ions depends mainly on (a) the presence of active spots due to isomorphous replacements and (b) their accessibility. In the kaolinite minerals where isomorphous replacement is negligible, base exchange properties are only weakly developed and are due mainly to broken edge effects. Whilst it is scarcely possible to form a quantitative estimate of the amount of a mineral present from its base-exchange capacity, methods based on this property may be invaluable for confirmatory purposes. Thus, a high base-exchange capacity suggests a montmorillonitic type of clay. Illites may show certain peculiarities, notably in their ability to fix potassium in a non-exchangeable form.

A distinction must be recognized between univalent and bivalent ions attached by electrostatic forces and freely exchangeable by other similar ions and those forming part of the lattice structure. Of these, the most important is magnesium, which enters the octahedral layer by substitution for aluminium. This magnesium is not ordinarily exchangeable, but may be rendered accessible and exchangeable by fine grinding which results in breakdown of the lattice. This is well illustrated by the investigations of W. P. Kelley, W. H. Dore, and S. M. Brown, who determined the content of exchangeable bases before and after grinding in a ball mill. Their results are shown in Table VIII.
### Table VIII.—Effect of Grinding Soil Colloids (Kelley, Dore, and Brown)

<table>
<thead>
<tr>
<th>Colloid</th>
<th>Milligram equivalents replaceable bases per cent.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Ca</td>
</tr>
<tr>
<td>Clay Adobe, Cal. ...</td>
<td>58.1</td>
</tr>
<tr>
<td>Clay Adobe, Cal. ground 30 hrs.</td>
<td>60.4</td>
</tr>
<tr>
<td>Clay loam, Cal. ...</td>
<td>54.7</td>
</tr>
<tr>
<td>Clay loam, Cal. ground 30 hrs.</td>
<td>59.9</td>
</tr>
<tr>
<td>Glacial drift soil, Ind.</td>
<td>54.0</td>
</tr>
<tr>
<td>Glacial drift soil, Ind. ground 72 hrs.</td>
<td>57.3</td>
</tr>
<tr>
<td>Limestone soil, Tenn.</td>
<td>32.8</td>
</tr>
<tr>
<td>Limestone soil, Tenn. ground 30 hrs.</td>
<td>41.7</td>
</tr>
<tr>
<td>Cecil Clay Subsoil, Ala.</td>
<td>14.2</td>
</tr>
<tr>
<td>Cecil Clay Subsoil, Ala. ground 30 hrs.</td>
<td>14.0</td>
</tr>
</tbody>
</table>

Whilst little effect was produced on the exchangeable calcium a considerable increase in exchangeable magnesium and potassium occurred and a smaller increase in exchangeable sodium. The increase in exchangeable magnesium might be ascribed to the breakdown of structures of the montmorillonite type and the increase of potassium to breakdown of illitic structures, for it is known that the latter types can fix potassium in non-exchangeable form. (For a fuller description of base exchange see Chapter VI.)

From the foregoing account of the development of ideas on the clay complex, it is evident that the indefinite absorption compound hypothesis (Van Bemmelen and Wiegner) which replaced the older kaolinite hypothesis must now give place to a conception of the clay complex as consisting of a few minerals of definitely crystalline structure. These minerals may not always exist separately but can occur as mixed structures. Nor must it be assumed that they are
wholly mineral in character, for organic groups may be associated in the lattice structure. About this association, however, comparatively little is known, apart from its possibility.

Although most of the clay complex consists of definitely crystalline material, amorphous material may also be present, particularly in soil horizons. Such material will be generally similar in composition to clay minerals in that it consists mainly of silica, alumina and ferric oxide, but will be associated with humus and represent the products of contemporary pedogenic processes.

Since the clay-humus complex includes replaceable cations, it is potentially of an acid character. These cations may be replaced by hydrogen and the base-desaturated complex thus obtained. There is some evidence that a limited replacement of anionic groups is possible. The exchange activity is partly due to interplanar ions held by electrostatic forces and partly to active spots at broken edges.

**SESQUIOXIDES IN THE CLAY COMPLEX**

As adventitious constituents in the clay complex, hydrated forms of alumina, notably gibbsite, ferric oxide as haematite, and hydrated ferric oxide as goethite or limonite may be present.

S. Mattson has carried out certain experiments which show that clay complexes low in silica contain their excess of sesquioxides as adventitious constituents. Colloidal clay from a number of red or yellow soils with low silica-sesquioxide ratios was treated with hot saturated aluminium chloride solution. In each case a grey residue having a silica-sesquioxide ratio approximating to 2.0 was obtained. This is the ratio demanded by the kaolinitic group of minerals, which have been shown to be characteristic minerals of highly weathered humid soils.

F. Hardy, using a method based on alizarin absorption for the identification and estimation of free sesquioxides in
soils, has found that, even in soils which contain a notable excess of sesquioxides as shown by their analyses, the proportions of free sesquioxides are generally small. For example, a lateritic clay from Cuba contained only 7.5% alumina and 8.3% ferric oxide, whilst red soils from Bermuda contained 11.2% alumina and 5.1% ferric oxide. Further, a considerable proportion of the free sesquioxides occurs in the coarser fractions of the soil. Hardy's data appear incompatible with the theory that the weathering complex of sesquioxide soils consists of a mineral of the general composition \( R_2O_3 \cdot 2SiO_2 \cdot xH_2O \) with all the excess sesquioxides present as such, and seem to imply the occurrence of compounds with less than two molecules of silica to each molecule of sesquioxide. It is, of course, possible that the sesquioxides in excess of the \( R_2O_3 \cdot 2SiO_2 \) formula may be lightly combined and extractable by the hot aluminium chloride solution used in Mattson's experiments.

Certain tropical desert soils, in spite of the restriction of leaching by arid conditions, may show bright red colours, suggesting the liberation of sesquioxides. It seems possible that in such cases the stability of the siliceous clay minerals is affected by high temperatures. The writer has found silica-sesquioxide ratios considerably greater than 2.0 for the clay of certain red sediments of the Trias, a formation which was deposited, in Britain, under desert conditions.

The iron compounds of the soil, although not occurring exclusively in the clay complex may be conveniently considered in connexion with our treatment of this subject, particularly as they are among the most reactive constituents of soils.

Iron is present, mainly as ferrous iron, in the unweathered minerals of the soil, particularly in the ferromagnesian minerals. Such iron is comparatively non-reactive, but passes into active forms principally oxides, by oxidation. Iron pyrites occur in many rocks, but rapidly undergoes oxidation with access of air. Ferrous sulphide also is rapidly oxidized and occurs only in soil horizons that
are constantly anaerobic. Iron can partially replace aluminium in the lattice of certain clay minerals. Nontronite is the ferric isomorph of beidellite.

Most of the active iron in soils occurs as ferric oxides, of which there are many varieties. The anhydrous form is haematite, $\text{Fe}_2\text{O}_3$. B. B. Polynov\textsuperscript{47} mentions the following hydrated forms:

1. Limonite, formerly considered to be a definite hydrate, $\text{Fe}_2\text{O}_3 \cdot 3\text{H}_2\text{O}$.
2. Göthite, $\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O}$.
3. Xanthosiderite, $\text{Fe}_2\text{O}_3 \cdot 2\text{H}_2\text{O}$.
4. Hydrogöthite, $3\text{Fe}_2\text{O}_3 \cdot 4\text{H}_2\text{O}$.
5. Turgite or hydrohaematite, $2\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O}$.

According to N. S. Kurnakov and E. J. Rode (cited by Polynov), the only true hydrate is göthite. All the existing forms are considered to be solid solutions of water either in göthite (limonite, etc.), or, in haematite (turgite, hydrohaematite, etc.).

In addition to oxides of iron occurring as such, it is probable that there are also present more active forms of iron combined in complexes with humus and, possibly, with silicic acid. Many methods have been proposed for extracting the active iron of soils, for example, those of O. Tamm\textsuperscript{48}, using acid ammonium oxalate, of M. Drosdoff and E. Truog\textsuperscript{49}, depending on reduction of ferric oxide with hydrogen sulphide, and of H. G. Dion\textsuperscript{50} in which reduction is effected by nascent hydrogen from the action of ammonium tartrate solution on aluminium. Certain of these methods may also attack the less active oxides such as göthite.

Alumina, as gibbsite (hydrargillite), $\text{Al}_2\text{O}_3 \cdot 3\text{H}_2\text{O}$ or diaspore, $\text{Al}_2\text{O}_3 \cdot \text{H}_2\text{O}$, or their more hydrated forms, occurs in ferrallitic and lateritic soils, but may also be present in small quantities in the illuvial horizons of podsolic soils. It may also occur in complexes, similar to the iron oxide extractable by Tamm's reagent.
GENERAL REVIEW OF THE CONSTITUTION OF THE CLAY COMPLEX.

The information at present available on the clay minerals is only of a preliminary character, and one of the principal tasks in the future is the survey of the clay minerals of the principal world groups and sub-groups of soils. In the meantime, the existing information on the composition of the clay complex of different soils permits certain broad distinctions to be drawn.

The clearest distinction is between clays with high silica-sesquioxide ratios and clays with moderate or low silica-sesquioxide ratios. H. Harrassowitz suggests the term *siallitic* for soils with clay fractions having medium or high silica-alumina ratios, and *allitic* for soils with clay fractions of markedly-aluminous character. This suggestion has gained a certain amount of acceptance and we may conveniently speak of *siallitic* clays for the former type and *allitic* or, preferably, *ferrallitic*, for the latter type. Any attempt to indicate a critical limit is necessarily provisional, but there is some justification for considering a molecular silica-sesquioxide ratio of 2.0 as significant. Soils whose clay fractions have silica-sesquioxide ratios greater than 2.0 are generally greyish or brownish-grey in colour in the absence of organic matter, and probably do not contain appreciable proportions of sesquioxides, apart from illuvial accessions. On the other hand, soils whose clay fractions have ratios less than 2.0 generally betoken by their brownish, yellowish, or reddish mineral colours the presence of free ferric hydroxide, and, by inference, aluminium hydroxide. Such soils show grey colours only under conditions favouring reductive processes.

We may, therefore, conceive a clay of the first class as consisting of a range of minerals in which the montmorillonite or illite type is the more predominant the more marked is its siliceous character. On the other hand, a clay of the second class may be considered to be made up mainly of minerals of the type of kaolinite together with adventitious
sesquioxides. There is the possibility that clay minerals of low silica-sesquioxide ratios may be pyrophyllites in which there has been extensive substitution of aluminium for silicon as suggested by Marshall (p. 131), but it would be very exceptional to find a mineral of this class with a SiO₂/R₂O₃ ratio of 2 or less.

Clays of the first class may be developed under a number of different conditions.

I. They may be formed by primary weathering from crystalline rocks under conditions which hinder the removal of silicic acid. This may occur under the following circumstances:—

(a) Leaching may be restricted owing to an excess of evaporation over rainfall in semi-arid and arid climates.

(b) Drainage may be impeded, as in ground-water soils, or in soils developed from parent materials with impervious strata, such as many of the boulder clays of Northern Europe.

(c) Removal of silicic acid may be inhibited by the prevalence of a high base status.

II. They may be formed by the precipitation of dissolved silicic acid concomitantly with suspended matter in estuarine, lacustrine, and marine sediments. Such clays are characteristic of soils of secondary weathering. The clay of immature soils derived from calcareous sediments appears, from the writer’s experience, to be of an exceptionally siliceous type. It will be remembered that Mattson has demonstrated the siliceous character of the isoelectric precipitates formed in the presence of bivalent ions.

III. They may be formed through the removal of associated sesquioxides from the clay complex, by leaching. This can take place in the following processes:—

(a) Podsolization under acid humus.
(b) Solotization, consequent on the hydrolysis of the sodium clay of alkaline soils and the removal of sodium hydroxide by leaching, whereby the clay complex becomes unstable, and loses sesquioxides. In this process, there may be an actual liberation of silica. (See p. 389.)

In clays of this class, it may be supposed that minerals of the siliceous montmorillonite-illite type are present as characteristic, if not dominant, ingredients. A complete survey of the different types of siliceous clays is, however, necessary before this can be safely assumed. It would be of particular interest to ascertain if the removal of sesquioxides in podsolization and solotization leads to the formation of new minerals or merely to liberation of silica.

Clays of the second class are formed under humid conditions with free removal of percolating waters in drainage. The consequent lowering of the base status leads to instability of the more siliceous complexes. Clays with silica-sesquioxide ratios of about 2.0 are common in humid temperate climates, but it is not certain whether this represents the stable product over a wide range of humid conditions or whether it is merely a stage in degradation to more sesquioxidic complexes. We know that under typical conditions primary weathering can give rise to products consisting almost entirely of sesquioxides. Given sufficient time, would primary weathering under humid temperate conditions proceed to complete desilicification? At present there is no clear evidence for or against. We must be careful, however, to distinguish weathering, which results in the formation of the parent material from the parent rock, from pedogenesis, which is confined to the soil horizons and which may result in the differentiation of horizons with clay complexes differing markedly from those of their corresponding parent materials. In studying the effect of climate and other factors on clay composition it is probably safer to confine attention to the parent material below the solum.
THE CLAY COMPLEX

We may also distinguish a type of clay which has been enriched by illuvial accumulation. The B horizons of podsolic profiles and the upper horizons of certain lateritic profiles are demonstrably of this type; but it is also possible that illuvial deposition of sesquioxides may occur to some extent in tropical red loams, red earths, and, if we accept the views of A. Reifenberg, in terra rossa.

Although the clays of the second class are less reactive than clays of the first class, such reactivity as they possess implies the presence of constituents containing exchangeable bivalent and univalent cations. The constituents are probably less siliceous than the minerals of the montmorillonite class, but still remain to be investigated.

SIGNIFICANCE OF SILICA-ALUMINA AND SILICA-SESQUIOXIDE RATIOS

During recent years much attention has been given to the composition of the clay fraction as one means of characterizing soils. Significant information is given by the molecular silica-alumina (SiO₂/Al₂O₃) or silica-sesquioxide (SiO₂/Al₂O₃ + Fe₂O₃) ratio. The writer is strongly of opinion that the latter ratio is preferable. The use of the silica-alumina ratio implies that the ferric oxide, known to be present in the clay fraction, is uncombined with silica, and present as adventitious ferric oxide. Such an assumption cannot be made, and we are bound to assume that the clay minerals include not only aluminosilicates but also ferrisilicates, and possibly ferrosilicates.

The silica-sesquioxide ratio, though of great value in distinguishing different types of clay, must not be assumed to give a complete specification of the material to which it relates. Two clays may have the same silica-sesquioxide ratio and even the same alumina-ferric oxide ratio and yet be markedly different in constitution owing to the possibility of isomorphous substitutions. Complete definition can only
be attained when it is possible to record all the minerals and
the relative proportions of each present.

In general, the silica-sesquioxide ratio of clays of humid
tropical climates tends to be low, indicating the tendency to
ferralitization, whilst that of clays of arid regions and of cold
regions tends to be high.

E. M. Crowther, from a review of the data of W. O.
Robinson and R. S. Holmes (loc. cit.), relating to the com-
position of the colloidal clay of representative American soils,
has attempted to assess the relative effect of humidity and
temperature on the silica-alumina ratio. There appears to be
a negative correlation of this ratio with rainfall, and a pos-
tive correlation with temperature. But no general relation-
ship capable of expression as a formula is likely to be
discoverable, even although the enquiry be restricted to soils
of primary weathering, unless it is possible to allow for the
effect of age and the possibility of erosion. In humid
tropical soils, the tendency is for soils to become more
sesquioxidic in character with age; but it is difficult to see how
this tendency can be given quantitative expression and intro-
duced as a term into a general formula. The complications
introduced by erosion, whereby illuvial horizons are laid bare
at the surface, are equally obstructive in the way of obtaining
a general relationship between silica-alumina or silica-
sesquioxide ratio, humidity, and temperature.

One climatic effect on the silica-sesquioxide ratio of clay
may be, however, a fruitful subject for investigation, and
that is the tendency of eluviation in modifying the composi-
tion of the clay complex in different horizons of the soil pro-
file. Whilst under cool humid conditions the tendency is
towards an impoverishment of the upper horizons in sesqui-
oxides, resulting in a clay complex relatively rich in silica,
under humid tropical conditions, the losses by eluviation fall
on the silicic acid, and the clay complex tends to be enriched
in sesquioxides.

Abundant examples of the decrease in silica-sesquioxide
ratio from the surface downwards are obtainable from pro-
file examinations in humid temperate and cool regions. Recently, L. D. Baver and G. D. Scarseth\textsuperscript{49} have studied the composition of colloidal clay from profiles in Alabama. Their results for soils derived from an unconsolidated parent material with a high silica-sesquioxide ratio plainly show an increase in the ratio from the surface downwards, betokening an eluviation of silicic acid. The northern limit of this type of weathering is placed at the 61°F. mean annual isotherm. Between this and the region of podsolic eluviation lies a transition zone, that of the brown earths or brown forest soils, in which there is no marked predominance of silicic acid or sesquioxides in the losses by eluviation, and in which, consequently, no marked variations in the silica-sesquioxide ratio of the clay complex are to be expected in passing down the profile. From such evidence as is obtainable it appears that in this region the tendency is for primary weathering to produce a type of clay complex whose silica-sesquioxide ratio approaches to 2.0. It would not be correct, however, to regard this as due to climatic factors alone. The effect of parent material must also be considered. In the same locality and under the same climate, acid rocks, such as quartzites, may yield podsolic profiles, whilst basic rocks, such as dolerites, may yield brown earths. The full implications of the distinction between the products of deep weathering and the actual soil horizons have also not been fully realized.

A. Reifenberg\textsuperscript{50}, from an exhaustive examination of published data on the composition of colloidal clay, has shown that the main soil groups have distinctive $\text{SiO}_2/\text{R}_2\text{O}_3$ ratios. He gives the following as average figures:

<table>
<thead>
<tr>
<th>Soil Type</th>
<th>$\text{SiO}_2/\text{R}_2\text{O}_3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grey desert soils</td>
<td>3.62</td>
</tr>
<tr>
<td>Red desert soils</td>
<td>2.08</td>
</tr>
<tr>
<td>Alkali soils</td>
<td>3.01</td>
</tr>
<tr>
<td>Prairie soils and chernozems</td>
<td>3.17</td>
</tr>
<tr>
<td>Podsols (A)</td>
<td>2.84</td>
</tr>
<tr>
<td>Terra rossa</td>
<td>2.43</td>
</tr>
<tr>
<td>Brown earths</td>
<td>1.98</td>
</tr>
<tr>
<td>Tropical red earths</td>
<td>1.73</td>
</tr>
<tr>
<td>Lateritic (ferrallitic) soils</td>
<td>1.28</td>
</tr>
</tbody>
</table>
SOILS

BIBLIOGRAPHY


CHAPTER VI.

BASE EXCHANGE AND OTHER REACTIONS OF THE COLLOIDAL COMPLEX

ABSORPTION

The colloidal complex is the seat of the most important chemical reactions of the soil, for, except in the case of calcareous, gypseous, and saline soils, the non-colloidal material is relatively inert, and chiefly significant from the physical standpoint.

One of the most characteristic properties of colloids is their ability to absorb dissolved substances from solution. This property is shown markedly by soils. It is, however, important to distinguish two types of absorption. The first type, more properly termed adsorption, depends on the change in interfacial tension produced by the presence of a dissolved substance in the liquid phase. Substances which lower interfacial tension tend to accumulate at an interface, whilst substances which raise interfacial tension show negative absorption. This type of absorption by the soil, though demonstrable, is of negligible moment in the pedogenic processes.

The second type of absorption, on the other hand, is of the highest importance in the study of the soil, and is due to reactions which are essentially chemical and ionic in character, but located at the surfaces of colloidal particles. The enormous development of specific surface implied by colloidal subdivision gives to these reactions their distinctive character. As usually understood, they are comprised under the term base exchange*, but it should be realised that these are not the only manifestations of surface chemical activity.

*The chemical purist may prefer the term cation exchange.
of which the soil colloidal complex is capable, and it may be possible to speak also of anion exchange. Whilst the reactions which are about to be discussed have been most intensively studied in respect of the mineral colloid or clay complex, it is more correct to regard them as the property of the colloidal complex as a whole, including both the mineral and the organic colloidal material.

**BASE EXCHANGE**

When an ordinary soil is allowed to react with the solution of a neutral salt, such as potassium chloride, a proportion of the potassium is absorbed or fixed by the soil and an approximately equivalent amount of calcium is displaced into solution. The reaction may be expressed as

\[ \text{Ca—soil} + 2\text{KCl} = \text{K}_2 — \text{soil} + \text{CaCl}_2 \]

Certain soils, known as acid soils, under similar circumstances, give up acid hydrogen to the solution, and the reaction may be expressed as

\[ \text{H—soil} + \text{KCl} = \text{K} — \text{soil} + \text{HCl} \]

These and related phenomena are described as base exchange reactions and have afforded a great body, of information as to the nature of the reactive constituents of soils. The cations which are concerned in base exchange are termed exchangeable or replaceable bases.

The modern study of the subject dates from the work of H. S. Thompson\(^1\) and J. T. Way\(^2\) in the middle of the nineteenth century. The stimulus to their work was provided by the fear that the newly introduced soluble fertilizers might be washed out of the soil in drainage. Thompson first showed that ammonium sulphate in solution undergoes a reaction with the soil whereby ammonium is fixed and calcium liberated. The careful and exhaustive investigations of Way demonstrated the chemical equivalence of absorption and displacement when soils react with solutions of neutral salts, and located the seat of these reactions in what we now know as the colloidal or clay-humus complex. It was also shown by Way that the speed of absorption was
very great and that, at a certain concentration of the dissolved salt, an upper limit of absorption was reached.

The absorptive properties of soils have been extensively studied since the discoveries of Way and Thompson, and many theories have been advanced to account for the observed facts. Liebig maintained that absorption by soils was a purely physical process analogous to the adsorption of dyes by charcoal. Whilst this type of absorption, depending on surface tension effects, may undoubtedly occur in soils, Liebig’s theory did not provide a satisfactory explanation of the reactions of soils with soluble salts. Way considered these reactions as essentially chemical, but the development of colloidal chemistry during the latter half of the last century proved an important help to the study of absorption reactions, and suggested the explanation that these phenomena, though chemical, are distinguished from ordinary chemical reactions by the fact that they are localized at surfaces.

We may now consider some of the characteristics of the reactions between soils and solutions of electrolytes. When a soil is allowed to react with a solution of an electrolyte such as potassium chloride, potassium is absorbed by the soil and an equivalent amount of other ions, principally calcium, is displaced into the solution. The chloride ion is not only unabsorbed, but may actually show negative absorption, whereby the chloride concentration in the equilibrium solution is greater than in the gel water associated with the soil colloids. Absorption is, however, incomplete; for, even with the most dilute solutions, a certain proportion of the absorbable ion remains in solution. G. Wiegner, from a review of available data, showed that the relationship between the concentration of an ion in the solid phase—the soil—and the concentration in the equilibrium solution could be expressed by the so-called Freundlich equation: \( y = K C^{1/p} \), where \( y \) is the amount in unit weight of soil, including that originally present, \( C \) the concentration of the ion in the solution, and \( K \) and \( p \) are constants.
Regarding, now, the ions liberated from the soil by a displacing ion, these will also be distributed between soil and solution, and displacement will be only partial. It is possible, however, by successive treatments of a quantity of soil with a salt solution, removing the equilibrium solution after each treatment, to obtain an almost complete displacement of the exchangeable bases originally present by the cation of the electrolyte used. This can be most readily effected by continuous leaching of the soil, as in the method of D. J. Hissink for the determination of exchangeable bases. The exchangeable bases in 25 g. of soil can be removed by leaching on a filter with one litre of normal sodium chloride solution. At the conclusion of this process, the exchangeable bases originally present in the soil have been replaced by an equivalent amount of sodium, and the soil may then be termed a sodium soil.

Hissink and others have shown that base exchange is practically instantaneous—a circumstance which is readily intelligible on the assumption that it is a surface reaction.

In Table IX, data are given for the content of exchangeable bases of some typical soils.

**Table IX. — Exchangeable Bases in Milligram Equivalents* in Typical Soils**

<table>
<thead>
<tr>
<th>SOILS</th>
<th>Ca</th>
<th>Mg</th>
<th>K</th>
<th>Na</th>
<th>Authority</th>
</tr>
</thead>
<tbody>
<tr>
<td>25 Dutch clay soils</td>
<td>30.0</td>
<td>5.0</td>
<td>0.8</td>
<td>2.5</td>
<td>D. J. Hissink (loc. cit.)</td>
</tr>
<tr>
<td>17 Scottish soils</td>
<td>9.95</td>
<td>0.776</td>
<td>0.242</td>
<td>0.269</td>
<td>A. M. Smith</td>
</tr>
<tr>
<td>7 Neutral soils, U.S.A.</td>
<td>13.92</td>
<td>4.83</td>
<td>0.75</td>
<td>1.48</td>
<td>W. P. Kelley and S. M. Brown</td>
</tr>
<tr>
<td>5 Alkali soils, U.S.A.</td>
<td>0.0</td>
<td>0.80</td>
<td>1.65</td>
<td>6.88</td>
<td>W. P. Kelley and S. M. Brown</td>
</tr>
<tr>
<td>6 Acid soils, U.S.A.</td>
<td>1.06</td>
<td>0.68</td>
<td>0.13</td>
<td>0.51</td>
<td>W. P. Kelley and S. M. Brown</td>
</tr>
</tbody>
</table>

*In the statement of results of exchangeable base determination it is convenient to use milligram equivalents rather than the actual percentages of exchangeable bases. A milligram equivalent is the equivalent weight in milligrams, so that 1 mg. equivalent of CaO = 0.028 g. A percentage of CaO is thus converted to mg. equivalents by dividing by 0.028. The mg. equivalent content of the other bases is computed in the same way.
It will be seen that, apart from alkaline soils, the principal base is calcium, accompanied by smaller amounts of magnesium, potassium, and sodium. Hissink found that, calculating all bases to milligram equivalents, the average composition of the exchangeable bases of 25 Dutch clay soils was

<table>
<thead>
<tr>
<th>Ca</th>
<th>Mg</th>
<th>K</th>
<th>Na</th>
</tr>
</thead>
<tbody>
<tr>
<td>79</td>
<td>13</td>
<td>2</td>
<td>6</td>
</tr>
</tbody>
</table>

Different figures are obtained for other types of soil, but, in normal soils, the four bases occur in proportions not widely different from those found by Hissink. In saline and alkaline soils, on the other hand, sodium is the predominant exchangeable base.

BASE-SATURATION AND UNSATURATION

Soils which have long been in equilibrium with calcium carbonate are said to be saturated with exchangeable bases. Under field conditions, the maximum capacity of a soil for exchangeable bases is the content of total bases when the soil is in equilibrium with excess of calcium carbonate. Any base present in excess of this amount appears in the soil in the form of carbonate.

Where calcium carbonate is absent, the content of exchangeable bases may fall below that representing saturation and the soil is said to be base-unsaturated or, simply, unsaturated. Such a condition may be produced under laboratory conditions by leaching a soil with dilute acids, such as hydrochloric or acetic acid, whereby the exchangeable bases go into solution as chlorides or acetates, and the soil becomes eventually base-free or desaturated. The gradual removal of bases occurs in nature, where soils, in the absence of free calcium carbonate, are subjected to the leaching action of percolating waters. The solvent action in such cases is augmented by the presence of dissolved carbon dioxide, which has originated from the respiration of plant roots and from the micro-biological decomposition of plant residues in the soil. The presence of sulphuric acid from the oxidation
of sulphur compounds also increases the rate of loss of bases by leaching. Alternate leaching and desiccation may result, according to H. G. Coles and C. G. T. Morison\textsuperscript{7} in increased removal of exchangeable bases. By repeatedly heating at 98°C., washing, drying, and re-heating, they were able in some cases to effect complete removal of bases, which went into solution mainly as sulphates.

In extreme cases, as for example, in very humid climates and with highly pervious soils, an almost complete removal of exchangeable bases, resulting in desaturation, may take place. For example, in the uplands of Wales, uncultivated soils frequently contain less than 0.01% of exchangeable calcium oxide. Very low figures may be encountered in light sandy soils.

Unsaturation may sometimes exist in soils which contain free calcium carbonate, notably in soils derived from hard limestones. The calcium carbonate may occur in relatively coarse hard fragments, and saturation is attained only in the immediate vicinity of such fragments, whilst the body of the soil is subjected to loss of exchangeable bases by leaching. Similarly, the remains of former dressings of lime or chalk may persist as coarse fragments, although the soil as a whole has become unsaturated.

The rate of loss will depend on the base-status and on the intensity of leaching. At Rothamsted the annual loss from soil containing free carbonate is 400-500 lbs. CaO per acre (cf. p. 36). In N. Wales, R. Williams\textsuperscript{8} found a loss of only 100-200 lbs. CaO per acre from soils with 0.2 to 0.4% exchangeable CaO. The higher rates of loss in N. Wales were found in those soils with the highest lime status. It may be presumed that in extremely acid soils the losses of lime are small and may be balanced by liberation from lime-bearing silicate minerals in the soil.

**EXCHANGEABLE BASE CONTENT AND SOIL REACTION**

The greater the degree of unsaturation of a soil, the more marked are its acid properties. The intensity of this acidity
Fig. 7.—Relationship between pH and exchangeable CaO of Anglesey soils.
is commonly measured by the reaction of a soil-water system, expressed as its pH value, i.e., the negative logarithm of its hydriion concentration. As the exchangeable base content of a soil falls, the reaction of its aqueous suspension becomes more acid. We shall discuss the precise significance of soil pH more critically at a later point. For the present we shall consider pH in its usual and conventional significance, namely, the pH of a soil suspension with a soil-water ratio of \(1:2.5\). Unless otherwise stated it can generally be assumed that soil pH values are on this conventional basis.

Since calcium accounts for about 80% of the reaction value of the exchangeable bases in soils, a close relationship should exist between the reaction, as shown by the pH value, and the exchangeable calcium, in soils with the same type of colloidal complex present in approximately the same proportions. This relationship is seen in Fig. 7, which shows graphically the relationship between pH and exchangeable calcium oxide for a number of soils of the Arvon series analysed for advisory purposes in the author's laboratory. These soils, derived from schistose drift under free-drainage conditions, contain about 17-20% of clay and about 5 to 8% of organic matter. The \(\text{SiO}_2/\text{R}_2\text{O}_3\) ratio of the clay is about 1.9.

It will be seen that there is a general tendency to a linear relationship. The "scatter" of the experimental points may be attributed to the varying proportions of organic matter and clay present and to small variations in the character of the clay itself.

Although, among soils of similar composition, there is an obvious relationship between the reaction of the soil, as expressed by its pH, and its content of exchangeable calcium oxide, a general expression for this relationship is not easily obtained, since the acidic properties of a soil vary not only with the relative proportions of mineral and organic matter in the colloidal complex, but also with the composition of the mineral absorbing complex. Whilst the pH of base-un-saturated organic matter may be 4.0 or even less, that of the
so-called clay acid is rather higher and depends mainly on
the relative proportions of acid (OH, SiO₂, P₂O₅, humic acid)
and basic (Al₂O₃, Fe₂O₃) groups present. Clay colloids rich
in silica, such as those of bentonite and certain alluvial clays
will be more acid when desaturated than clay colloids rich in
sesquioxides, such as those of ferrallites and tropical red
earths. (See also p. 184.)

The reaction of the soil suspension will also be affected,
independently of the base status of the colloidal complex, by
the presence of mineral acids. B. Heimath⁹ considers that
the pH in acid forest humus is mainly governed by the
sulphuric acid present in the soil moisture. The very low
pH values found by H. C. Doyne and R. R. Glanville¹⁰ for
certain Nigerian swamp soils are attributable to mineral
acids. The former author⁶⁴ found the pH of freshly cleared
mangrove soil to be 5.0-6.0. On being kept moist the pH
fell to 1.2.5, with formation of ferric hydroxide and free sul-
phuric acid from the oxidation of ferrous sulphide. After
washing away the excess of sulphuric acid the pH rose to
3.6-4.3. (See also p. 367.) In N. Wales, the writer found the
A horizon of a podsol developed from weathered Millstone
Grit with pH 3.28 to contain 0.098% free sulphuric acid.

DEGREE OF UNSATURATION AND EXCHANGE CAPACITY

The content of exchangeable bases at saturation, ex-
pressed in milligram equivalents, is termed the exchange
capacity. Much ingenuity has been shown in devising
methods for determining the extent of unsaturation of acid
soils, i.e., the amount of base required to produce saturation.
The difficulty consists in arriving at a satisfactory definition
of saturation. For, even when a soil contains such a propor-
tion of exchangeable bases that its reaction is neutral
(pH = 7.0), it can still combine with more bases. The case is
parallel with that of a weak acid such as acetic acid. If a
strong alkali such as soda be added to a given amount of
acetic acid in dilute solution, the quantity of soda added
when the neutral point is reached is less than the equivalent
of acetic acid present. When the soda is exactly equivalent to the acetic acid the reaction of the solution is markedly alkaline owing to hydrolytic dissociation.

D. J. Hissink\textsuperscript{11} has sought to measure the degree of unsaturation of soils by determining the amount of base with which a soil can combine in the presence of such an excess of alkali that hydrolytic dissociation is suppressed. Under ordinary conditions, however, a soil would never arrive at such a state of base-saturation, and it has therefore been proposed by various workers to determine the amount of base required to bring a soil to the point of neutrality. This can readily be ascertained by electrometric titration, by adding increasing amounts of a standard solution of alkali to a soil suspension, and determining the pH potentiometrically for each addition. There are also a number of methods in which the soil is leached with neutral salt solutions. For example, in the method of F. W. Parker\textsuperscript{12}, soil is leached with barium acetate, adjusted to neutrality, to obtain the barium soil at pH 7.0. The combined barium is then displaced by leaching with N. ammonium chloride solution and the absorbed ammonium determined by distillation. In the method of W. P. Kelley\textsuperscript{13} barium hydroxide is used to obtain barium saturation, and the sorbed barium determined after leaching with ammonium acetate. In each case the number of milligram equivalents of base present in the soil saturated at pH 7.0 is taken as the measure of the base-exchange capacity. In another group of methods exchangeable hydrogen is determined and this, together with the other bases present, is taken as the base exchange capacity. The problem is complex because hydrogen is strongly absorbed and replaced with difficulty by other cations. There is also a qualitative difference between the bonds in the 1:1 lattice type and those in the 2:1 type. In the former case there is little isomorphous replacement, and exchangeable hydrogen is probably bound to the oxygen of OH groups, probably at broken edges. In the 2:1 clays hydrogen is mainly held by electrostatic forces. Thus clays of this group are more dissociated
and tend to give lower pH values than clays of the 1:1 group.

In another class of methods, soil is allowed to react with calcium carbonate. The carbonate decomposed, measured by the amount of carbon dioxide evolved, is a measure of the degree of unsaturation and, if equilibrium is reached, actually represents, when the base equivalent is added to the equivalent of the bases originally present, the exchange capacity. The chief drawback to these methods is the difficulty of fixing an end-point to the reaction under laboratory conditions.

The question of base-unsaturation is of considerable importance from the practical standpoint, since, by means of it, an estimate is sought of the amount of lime or calcium carbonate necessary as a dressing in order to bring the soil to an optimum base-status. The difficulty of the problem lies in discovering what is the optimum base-status, or even the minimum base-status consistent with fertility, since both must vary according to climate and cropping. Whilst for certain crops, such as potatoes, a base-status definitely below saturation, corresponding with a markedly acid reaction, may be sufficient for satisfactory growth, other crops, such as lucerne, appear to require a higher base-status. The actual relationship between pH and plant growth has been studied by many workers. Whilst it may be generally stated that plant growth is adversely affected by very acid or very alkaline reactions (pH < 5.0 or > 8.5), individual crops show wide and ill-defined optima. Actually, the relationship is not simply between plant growth and soil reaction, for calcium supply and the possible effect of toxic aluminium in acid soils enter into the problem. (See p. 541.) From the plant physiological viewpoint, then, it is not yet certain whether the important factor is the soil reaction, its content of exchangeable bases, or its deficit from base-saturation. Actually, all these effects are linked. A low content of exchangeable bases implies an acid reaction and the presence of available aluminium.
THE SIGNIFICANCE OF SOIL pH VALUES

As was stated above, by soil pH is commonly understood the pH of a 1:2.5 soil-water suspension. It should be clearly realized, however, that soil pH thus understood is to a large extent of merely conventional significance and not an absolute value, such as, for example, the content of total nitrogen, total phosphorus, or exchangeable calcium.

The determination of the hydrogen-ion concentration or pH of a liquid or of a suspension is primarily a problem for the physical chemist, and for our purpose we may assume that the method used is capable of giving a true value. The uncertainty enters when it is necessary to decide as to the actual liquid or suspension on which the determination shall be made. How great this uncertainty is may be appreciated when it is realized that the pH determined on a water suspension of a particular soil may be affected by (1) the ratio of soil to water; (2) the carbon dioxide concentration of the suspension; and (3) the time of sampling.

(1) The conventional character of the determination of pH as usually practised is evident from the selection of the 1:2.5 ratio of soil to water, a selection based mainly on practical convenience and lacking any theoretical justification. As a general rule, the pH found falls as the ratio of water to soil is decreased. This has been shown by many investigators. For example, M. R. Huberty and A. R. C. Haas\(^{14}\) found that a range of citrus soils giving pH values about 7.0 with the conventional soil water ratio, gave figures approaching or below neutrality at low moisture contents. The latter author considered that few, if any, Californian orchard soils were alkaline at low moisture contents. L. E. Davis\(^{15}\), whilst admitting the decrease in pH with moisture content, throws doubt on some of the values obtained at very low moisture contents, which he considers to be due to electrode effects. Assuming, however, that such effects can be avoided the determination of pH at low moisture contents is likely to give a better insight into actual conditions within the soil than its determination in a dilute suspension. There
is no difficulty in carrying out the determination at the flow point (upper limit of plasticity). By determining the pH at other soil water ratios and extrapolating it might be possible to form an estimate of the probable pH at still lower soil water ratios.

(2) Small changes in CO₂ concentration can cause considerable changes in pH value. Within the soil the CO₂ concentration can vary considerably with general biological activity and is also affected by changes in the moisture-air regime. R. S. Whitney and R. Gardner¹⁶ have shown that the pH of a soil suspension with constant soil-water ratio is approximately a linear function of the carbon dioxide concentration from 0.0003 up to 1 atmosphere CO₂. They consider that the expression of pH as a function of CO₂ pressure would give a better indication of field conditions than would be obtained by a single determination. In any case the importance of controlling CO₂ partial pressure in pH determinations is evident.

(3) However carefully the determination of soil pH is made, there is still an uncertainty owing to seasonal variations. These may amount to as much as a unit of pH in exceptional cases, whilst variations of 0.5 are common. Generally speaking, the pH of acid soils will be highest after long continued wet weather and will fall during a dry period, but this is not invariably the case. Acidity due to mineral acids will be reduced by the washing action of rain. A number of soil processes, including nitrification, may be expected to influence the pH.

The production of carbon dioxide by micro-biological decompositions and by respiration of plant roots would tend to lower pH. A similar effect would be produced by the absorption of cations by plants. On the other hand, the absorption of anions, e.g., nitrate might be expected to lead to a rise in pH. The general effect of leaching in drainage is to lower pH, but this would be a continuous effect and seasonal only in so far as leaching is more intense in the winter than in the summer months.
R. K. Schofield\textsuperscript{17} has shown that variations in the concentrations of cations of the soil solution in equilibrium with the exchangeable cations of the soil can occur provided that the concentrations of all univalent ions are changed in one ratio, those of all the bivalent ions in the square of that ratio, and those of all the trivalent ions in the cube of that ratio. The consequence of this ratio law may be appreciated if we take the simple case where the only ions present are calcium and hydrogen; then, assuming the calcium-ion concentrations to be reduced to one quarter, equilibrium with the soil exchange complex would be maintained by a corresponding reduction of the hydron concentration to one half, i.e., a rise of 0.3 in the pH.

Although from the above considerations it is clear that soil pH as ordinarily determined is lacking in precision of meaning, the influence of hydrogen-ion concentration on most chemical and biological reactions in liquid media is so marked that it is of the highest importance to obtain a more precise definition. The pH that actually controls the reactions with the soil is that of the soil solution under ordinary field conditions. Most of this moisture is present in thin films in channels and interstices, conditions under which ionic equilibria differ markedly from those obtaining in bulk liquids as they are usually known to us. Moreover, on account of the heterogeneity of the soil mass and even of the colloidal surfaces, pH values must vary considerably at different points, even over small ranges, within the same soil. Indeed, it would appear that no single figure can adequately represent the pH pattern within an actual soil.

To some extent, then, the pH of the soil at any given moment must be a conventional figure. We have already referred to the uncertainty introduced by variations in soil-water ratio and in carbon di-oxide concentration. Various proposals have been made for defining an absolute pH. A. N. Puri and A. Sarup\textsuperscript{18} proposed the iso-hydric pH as such an absolute value. This is defined as the pH of a buffer solution that shows no change in reaction after contact with
the soil. S. Mattson\textsuperscript{19} similarly proposed the point of exchange neutrality, defined as the pH at which the addition of a neutral salt has no effect on the reaction of the soil. This is obtained by titrating a suspension of 10 g. soil in 20 ml. water and another of the same amount of soil in 20 ml. of N. sodium sulphate solution with 0.02 N alkali or acid, according as the soil is originally acid or alkaline. The reaction at the point of intersection of the two curves gives the point of exchange neutrality. W. T. McGeorge\textsuperscript{20} has shown that for a number of saline and alkaline Arizona soils these two definitions of pH are equivalent and that nearly the same value is given by measuring the pH of a soil paste of water content approximating to that of the moisture equivalent.

**BUFFERING CURVES AND BUFFER CAPACITY**

If varying amounts of acid or alkali are added to a given suspension of soil in water and the pH determined after equilibrium has been attained, it is possible to obtain so-called buffering curves. This may be exemplified by some data obtained by W. Brenner\textsuperscript{21}. Ten-gram portions of soil were shaken in flasks, each containing 30 c.c. of water, to which varying amounts of 0.1 N-hydrochloric acid and 0.1 N-calcium hydroxide, respectively, had been added. After allowing to stand for 24 hours with frequent shaking, the pH was determined electrometrically. Results for typical soils are shown in Fig. 8. The curve NN represents a blank experiment without soil. The divergence of an individual buffering curve from this curve gives a measure of the amount of interaction between the soil and the acid or alkali. It will be seen that whilst the curve for siliceous sand approaches closely to that for the blank, indicating only slight interaction, the curves for clay and raw humus show considerable divergence from this. In the curve for sand, it will be seen that, whilst 2 c.c. of acid cause a change of about 2.75 in pH, 2 c.c. of alkali cause a change of about 2.0 in pH. The sand is said to be more strongly buffered on the
Fig. 8.—Buffering curves of soils. (W. Brenner.)
alkaline than on the acid side. Comparing the change in pH produced by the addition of 5 c.c. of alkali, we have for the sand 4.5, for the loamy clay 1.3, for the podsol A horizon 1.0, and for raw humus 0.2.

If we measure the buffering capacity by the amount of alkali required to produce a given change in pH, the order given by the curves would be raw humus > podsol A-horizon > loamy clay > sand. In general we may say that the buffering capacity of soils depends on their content of reactive colloidal material. It is to be noticed that soils show buffering both on the acid and alkaline side. They are said to be amphoteric, i.e., exhibiting both basic and acidic properties. This is to be expected since the colloidal complex contains both acidoid and basoid components.

ENERGY OF ABSORPTION OF DIFFERENT CATIONS

The energy of absorption of cations by the colloidal complex varies considerably. The order of absorption of the principal ions is \( H > Ca > Mg > K > Na \). The replacement of exchangeable sodium by exchangeable hydrogen is, thus, more rapid than the replacement of exchangeable hydrogen by exchangeable sodium. The relative absorption of calcium and sodium is illustrated by certain data obtained in the Sudan Government Laboratories\(^{22}\). Sodium-saturated clay was repeatedly treated with solutions containing calcium and sodium chlorides which were decinormal with respect to chloride, but contained varying equivalent ratios of calcium to sodium. The ratios of calcium to sodium in the clay were determined after each treatment. The following is a selection from the results obtained:

| Ca/Na in solution | 2.33 | 1.50 | 1.00 | 0.8 | 0.42 | 0.20 | 0.11 | 0.058 |
| Ca/Na in clay | 14.2 | 10.6 | 9.3 | 6.8 | 4.9 | 2.66 | 1.59 | 0.76 |

It is seen that the clay in equilibrium with the solution containing equal equivalent amounts of calcium and sodium contained 9.3 times as much calcium as sodium, and that in order to obtain an excess of sodium in the soil the ratio of
sodium to calcium in the percolating solution had to be raised to the order of 15:1.

The relative energy with which bases enter the absorbing complex is also shown by some results obtained by A. F. Joseph and H. B. Oakley\textsuperscript{23}, who allowed equimolecular solutions of calcium, potassium, and sodium chlorides, of strength 0.5 normal with respect to chloride to percolate through a soil. With calcium and potassium chlorides, the ratio Ca/K attained in the soil was 0.9; with calcium and sodium chlorides the Ca/Na ratio was 6; and with potassium and sodium chlorides the K/Na ratio was 6. It appears, therefore, that the energy of replacement of calcium is slightly less than that of potassium and that both greatly exceed sodium in this respect.

The mutual replacement of cations associated with colloidal particles is connected according to G. Wieger\textsuperscript{24} and his co-workers with their ionic hydration and hence with their ionic radii.

The great energy of absorption of hydrogen is probably due to the fact that whilst the metallic cations form electrovalent bonds, hydrogen forms co-valent bonds. And, therefore, in order to obtain exchange of hydrogen ions, it is first necessary to secure their dissociation by the rupture of the co-valent bond.

A further factor in the relative absorption and replacement of cations is the possibility of the displaced cations forming undissociated or insoluble products. Although calcium is strongly absorbed, it will be readily replaced if it can form insoluble compounds, with oxalate or phosphate ions, for example.

The great energy of absorption of hydrogen, however it may be explained, renders it extremely difficult to determine the content of exchangeable hydrogen in a soil by leaching with salt solution. Nevertheless, a certain amount of replacement does occur, giving rise to what is termed exchange acidity. Methods have been proposed for the determination of exchange acidity and hence of the base-acid
status of soils, but it should be realized that they do not give the total content of exchangeable hydrogen. Methods for determining the degree of unsaturation of soils from the pH of their suspensions in electrolytes are purely empirical.

It should be added that considerable difference of opinion exists as to the nature of the reaction whereby exchange acidity is developed. Whilst it is usual to consider it as a simple replacement of exchangeable hydrogen, many workers consider that aluminium is actually replaced and that the acidity is due to hydrolysis of the resultant aluminium salt. This explanation is probable in the case of kaolin and ceramic clays. (See p. 185.)

FACTORS AFFECTING THE EXCHANGE CAPACITY

It has been already stated that the property of base exchange in soils resides in the colloidal complex, which is composed of clay and humus. The exchange capacity will therefore depend on the proportions of each of these constituents present in the soil and also on their individual exchange capacities.

Varying figures are obtained for the exchange capacities of clay and humus according to the nature and origin of the materials, and the methods employed for determining exchange capacity. For colloidal clay, the figures obtained by S. Mattson, using the method of leaching to equilibrium with neutral calcium chloride solution, range from 0.164 to 1.102 mg. equivalents per gram, the lowest figure being given by ferrallitic clay and the highest by bentonite. Marked correlation between exchange capacity and silica-sesquioxide ratio is shown. Estimates for the exchange capacity of humus vary from about 2.5 to 4.5 mg. equivalents per gram.

Considering saturation to be represented by the base-status of soils which have long been in equilibrium with excess of calcium carbonate, R. Williams has sought to determine the relative base-binding capacity of clay and organic matter by comparing the total exchangeable base
content of a number of calcareous soils with their content of clay and organic matter. The data are shown in Table X.

**Table X.—Relation between Total Exchangeable Bases, Clay Content, and Organic Carbon Content for Carbonate Soils**

<table>
<thead>
<tr>
<th>SOIL</th>
<th>Clay %</th>
<th>SiO₂</th>
<th>R₂O₅ of clay</th>
<th>Organic Carbon Total</th>
<th>Oxidizable</th>
<th>CaCO₃ %</th>
<th>Total bases in mg. equivalents calculated by Formula 1</th>
<th>Formula 2</th>
<th>Total bases in mg. equivalents %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vaynol Clay...</td>
<td>63.5</td>
<td>2.33</td>
<td>0.64</td>
<td>0.14</td>
<td>13.6</td>
<td></td>
<td>39.1</td>
<td>37.1</td>
<td>36.8</td>
</tr>
<tr>
<td>B 1459</td>
<td>54.0</td>
<td>2.34</td>
<td>1.54</td>
<td>0.06</td>
<td>8.3</td>
<td></td>
<td>37.8</td>
<td>35.8</td>
<td>36.7</td>
</tr>
<tr>
<td>G. 109</td>
<td>41.7</td>
<td>2.73</td>
<td>4.45</td>
<td>2.43</td>
<td>7.0</td>
<td></td>
<td>44.0</td>
<td>40.1</td>
<td>40.0</td>
</tr>
<tr>
<td>Harpenden</td>
<td>24.7</td>
<td>2.01</td>
<td>1.66</td>
<td>3.59</td>
<td>4.7</td>
<td></td>
<td>18.8</td>
<td>17.8</td>
<td>18.6</td>
</tr>
<tr>
<td>G. 26 A</td>
<td>39.3</td>
<td>4.78</td>
<td>2.54</td>
<td>2.06</td>
<td>3.6</td>
<td></td>
<td>34.1</td>
<td>35.4</td>
<td>34.4</td>
</tr>
<tr>
<td>G. 115 A</td>
<td>27.5</td>
<td>2.58</td>
<td>1.94</td>
<td>4.45</td>
<td>3.3</td>
<td></td>
<td>24.4</td>
<td>24.8</td>
<td>25.9</td>
</tr>
<tr>
<td>G. 150 A</td>
<td>34.3</td>
<td>2.84</td>
<td>3.14</td>
<td>3.22</td>
<td>3.1</td>
<td></td>
<td>32.8</td>
<td>33.3</td>
<td>33.0</td>
</tr>
<tr>
<td>Llyfasli</td>
<td>27.0</td>
<td>3.79</td>
<td>2.29</td>
<td>1.83</td>
<td>3.1</td>
<td></td>
<td>25.8</td>
<td>26.9</td>
<td>25.6</td>
</tr>
<tr>
<td>Rendzina</td>
<td>27.0</td>
<td>3.79</td>
<td>1.63</td>
<td>1.39</td>
<td>2.7</td>
<td></td>
<td>41.4</td>
<td>42.2</td>
<td>40.4</td>
</tr>
<tr>
<td>Groes Morrid</td>
<td>9.3</td>
<td>2.01</td>
<td>1.89</td>
<td>1.38</td>
<td>2.6</td>
<td></td>
<td>43.9</td>
<td>45.3</td>
<td>43.7</td>
</tr>
<tr>
<td>F. 85</td>
<td>7.8</td>
<td>2.01</td>
<td>2.00</td>
<td>1.39</td>
<td>2.4</td>
<td></td>
<td>28.5</td>
<td>29.7</td>
<td>28.5</td>
</tr>
<tr>
<td>G. 117</td>
<td>21.2</td>
<td>1.79</td>
<td>2.30</td>
<td>1.01</td>
<td>2.2</td>
<td></td>
<td>23.0</td>
<td>24.2</td>
<td>23.4</td>
</tr>
<tr>
<td>D. 129</td>
<td>33.6</td>
<td>1.69</td>
<td>3.34</td>
<td>2.62</td>
<td>1.7</td>
<td></td>
<td>34.3</td>
<td>35.6</td>
<td>33.9</td>
</tr>
<tr>
<td>G. 206</td>
<td>23.6</td>
<td>1.99</td>
<td>7.21</td>
<td>3.07</td>
<td>8.0</td>
<td></td>
<td>46.25</td>
<td>42.8</td>
<td>45.7</td>
</tr>
<tr>
<td>G. 178 A</td>
<td>47.1</td>
<td>n.d.</td>
<td>3.62</td>
<td>2.95</td>
<td>3.3</td>
<td></td>
<td>43.3</td>
<td>45.4</td>
<td>38.7</td>
</tr>
<tr>
<td>G. 57....</td>
<td>22.9</td>
<td>n.d.</td>
<td>2.98</td>
<td>1.84</td>
<td>0.8</td>
<td></td>
<td>26.6</td>
<td>24.6</td>
<td>13.6</td>
</tr>
<tr>
<td>F. III....</td>
<td>15.0</td>
<td>n.d.</td>
<td>3.28</td>
<td>2.20</td>
<td>0.5</td>
<td></td>
<td>25.8</td>
<td>24.7</td>
<td>19.3</td>
</tr>
<tr>
<td>Ab 3-5....</td>
<td>22.4</td>
<td>n.d.</td>
<td>3.17</td>
<td>2.36</td>
<td>0.2</td>
<td></td>
<td>27.2</td>
<td>27.6</td>
<td>14.5</td>
</tr>
</tbody>
</table>

Organic carbon has been used as a measure of the humus and two figures have been obtained in each case, namely, total carbon, and organic carbon removable by prolonged boiling with 4% hydrogen peroxide. The latter determination was made in order to avoid the error in total organic carbon introduced, in some cases, by the presence of the fragments of coal and cinder which are so frequent in cultivated soils.

By the use of a graphical method, equations were obtained for the relationship between total bases (T), clay (K), total carbon (C₁), and oxidizable carbon (C₀). The relationships thus found are:

\[ T = 0.57 \times K + 4.55C₁ \]  \hspace{1cm} (1)

\[ T = 0.57 \times K + 6.3C₀ \]  \hspace{1cm} (2)

The calculated values are shown in columns 6 and 7 respectively.
BASE EXCHANGE

It will be seen that the agreement between the calculated values for total bases and those actually found is fairly satisfactory for the first thirteen soils. G. 206 contains a considerable proportion of coal fragments, which leads to a high figure being given by Formula 1. On the other hand, the figure by Formula 2 is rather low.

The last four soils all show experimental values markedly below calculated values. In the case of G. 178 A, most of the carbonate is present as coarse hard fragments, whilst in the case of the remaining three soils the small amounts of carbonate present probably represent fragments surviving from calcareous dressings. The result is that these soils are definitely unsaturated.

The exchange capacity indicated for clay, namely, 0.57 mg. equivalents per gram falls within the limits found by S. Mattson (loc. cit.), and is in good agreement with a figure obtained by E. M. Crowther and J. K. Basu for Woburn soils.

Assuming soil organic matter to contain 58% of carbon, the exchange capacities are 2.64 mg. equivalents per gram of total organic matter and 3.65 mg. equivalents per gram of oxidizable organic matter, respectively. Both figures fall within the limits given at the beginning of this section.

Williams' results give no evidence for a correlation between exchange capacity and the SiO\textsubscript{2}/R\textsubscript{2}O\textsubscript{3} ratio of the clay. This is in apparent conflict with the findings of Mattson mentioned above; but it must be remembered that these results refer to exchange capacity as determined on soils which have been for long periods in the presence of excess of calcium carbonate. Further, none of the soils was markedly sesquioxidic. It may be that under such conditions a modification of the exchanging complex has taken place. Indeed, according to Mattson’s iso-electric theory, the clay complex should undergo modification in the direction of that type of complex which is stable at the prevailing pH. Although some of the clay analyses show comparatively low SiO\textsubscript{2}/R\textsubscript{2}O\textsubscript{3} ratios, it may be that in such cases there are pre-
sent small quantities of material of high exchange capacity and wide SiO₂/R₂O₆ ratio formed by modification of the original complex. Under the conditions of laboratory experiment, time would probably not suffice for such new formation.

**EFFECT OF EXCHANGEABLE BASES ON SOIL PROPERTIES**

The nature and content of exchangeable bases in a soil have an important bearing on its general properties and behaviour.

Soils with high base-status in which the principal saturating cation is calcium are generally most satisfactory from the agricultural standpoint. Such soils readily assume a granular structure, which minimizes the unfavourable effects of a high clay content and ensures good aeration and free percolation of water received as rainfall or irrigation. The clay complex tends to be of a siliceous type and there is no differentiation of the soil into horizons with clay of differing composition. The humus is also stable against leaching. Soils of this class occur generally in arid regions where leaching is incomplete owing to excess of evaporation over rainfall. In humid regions, soils containing natural calcium carbonate (rendzinas) or soils to which large calcareous dressings have been given belong to this class. In general, it may be said that one of the aims of cultivation and manuring should be to produce and maintain a calcium soil.

The pH of calcium-saturated soils having an excess of calcium carbonate may run up to over 8.0. In the absence of excess calcium carbonate the pH will be in the region of 7.0. As the exchangeable calcium falls the pH falls until, with complete unsaturation, a pH figure of about 4.0 is reached. A soil which is completely or almost completely unsaturated may be termed a hydrogen soil. Its actual pH will depend on the character of the colloidal material (see p. 184) and on the presence of mineral acids such as sulphuric acid. Unsaturation is associated with instability of the colloidal com-
plex. This instability is of two types. Firstly, there is physical instability, the tendency for the colloid to become dispersed or peptized. This is exhibited most markedly when the colloid is rich in silicic acid or humus. With more sesquioxodic complexes the tendency to dispersion in unsaturated soils is less marked. The dispersion of the complex favours the development of clay or humic B horizons. Secondly, there is chemical instability, whereby the clay complex itself undergoes decomposition. Here we may distinguish two possibilities. In one case, where humic acids are present there is a mobilization and leaching of sesquioxides (podsolization). In the other case, where, owing to the rapid mineralization of organic material, humic acids are absent, it is the silicic acid which becomes mobile and the residual material tends to become more sesquioxodic (ferrallitization). An intermediate case is that where there is a partial removal of silicic acid and a stage is reached in which any removal of silicic acid by leaching is balanced by a corresponding loss of sesquioxides (brown earth formation). Podsolization and ferrallitization mark the two extremes of decomposition of the clay complex consequent on desaturation.

A special case is presented by soils in which the dominant cation is sodium, and it should be noted that in order for the characteristic properties of sodium soils to be shown, it is not necessary that the sodium ion should be in actual excess. It has long been known that the presence of sodium salts in the soil is attended by marked changes in physical properties. These were attributed to the formation of sodium carbonate. The reaction, as conceived by Hilgard, was represented by:

$$\text{Na}_2\text{SO}_4 + \text{CaCO}_3 = \text{Na}_2\text{CO}_3 + \text{CaSO}_4,$$

assuming sodium sulphate to be the salt originally present. This theory has been proved untenable and the reactions involved have now been satisfactorily explained by the investigations of A. De Domenicis, K. Gedroiz, A. A. J. De 'Sigmond, and A. B. Cummins and W. P. Kelley.
If a quantity of a base-saturated calcium soil is treated with sodium chloride solution, calcium will be displaced and the sodium soil produced. When the excess of sodium chloride is removed by leaching the sodium soil undergoes a partial hydrolysis which may be expressed by

\[ \text{Na-soil} + \text{H}_2\text{O} = \text{H-soil} + \text{NaOH} \]

The deflocculation of the sodium clay and the higher pH of its aqueous suspension compared with that of the calcium clay is due primarily to the greater dissociation of the sodium compared with the calcium clay. With the removal of excess sodium chloride the sodium clay undergoes partial dissociation and the sodium humate present goes into solution, giving dark coloured solutions. Under natural conditions, small amounts of sodium carbonate are formed either by the carbonation of sodium hydroxide, or by a direct reaction of the sodium soil with calcium carbonate, if present. Sodium carbonate is not formed by direct reaction of a soluble salt with calcium carbonate.

A fuller discussion of sodium soils is given in Chapter XIII.

PROPERTIES OF SUSPENSIONS

We have seen that the colloidal material of the soil is built up from (a) acidoid groups, principally silica, and (b) basoid groups, principally alumina and ferric oxide. The properties of a colloidal particle suspended in a liquid are governed primarily by the character of the outer layer, which, so far as soil colloids are concerned, is predominantly of an acidoid or anionic character. With the acidoid groups are associated corresponding cations, the so-called exchangeable cations.

The structure of a colloidal particle suspended in water is shown diagrammatically in Fig. 9.* The inner shell of

*The spherical arrangement is shown merely for convenience of representation and should not be taken to mean that the colloidal material consists of spherical particles. Actually, the exchangeable cations, so far as the clay complex is concerned, are associated with fragments of laminated structures, and in the case of the pyrophyllite group of clays are associated with exchange spots between the lattice planes.
anions and the outer shell of cations form the so-called double layer. Whilst the anionic groups form the outer layers of the colloidal particle, the corresponding cations arrange themselves with respect to the particle and the surrounding liquid in a distribution depending on their own

![Diagram of a colloidal particle with anions and cations of the double layer.](image)

*Fig. 9.*—Diagram of a colloidal particle with anions and cations of the double layer.
ionic activity, and their degree of hydration. The distribution is also governed by the presence of electrolytes according to the well known Donnan equilibrium, whereby the product of the concentrations of an ionic pair must be the same in all parts of the system.

The net result is that the cationic sheath forms a diffuse layer rather than a definite shell of cations all at the same distance from their corresponding anions. Some of the cations will form undissociated combinations with the sheath anions. Indeed, it is probable that only a small proportion of the exchangeable cations are actually dissociated and capable of contributing to the diffuse double layer. For example, S. Mattson has calculated that, in the case of a colloidal particle of radius 45.5 \( \mu \) with 502,208 exchangeable cations, a dissociation of only 857 ions would account for the observed electrical potential.

Since a proportion of cations is relatively remote from the particle, there is an excess of negative charges. In other words, the particle is negatively charged. The magnitude of the charge will be governed by the character of the colloidal surface and by the extent to which dissociation has taken place. The potential difference (P.D.) between the inner and the outer sheath is given by

\[ P.D. = \frac{ed}{Dr^2}, \]

where \( e \) is the charge on the particle, \( d \) the mean radial distance between the anionic sheath and the cationic sheath, \( D \) the dielectric constant, and \( r \) the radius of the particle. For a given particle, since \( e/Dr^2 \) is approximately constant, the potential will depend mainly on \( d \), the radial distance between the two sheaths. This is governed by the degree of hydration of the cations present and by their degree of dissociation. The more heavily hydrated these are, the greater is the distance between them and the inner layer. The order of hydration of the univalent cations

\[ *1 \ \mu = 0.00001 \ \text{mm}. \]
is $\text{Li} > \text{Na} > \text{K} > \text{Rb} > \text{Cs} > \text{H}$, and this is also the order for the potential of the corresponding particles of clay. The peculiar position of hydrogen as an exchangeable ion has already been noted. Although the degree of hydration of the hydrogen ions is somewhat uncertain, hydrogen clay particles show the lowest potential and the lowest stability.

It has been stated in the discussion of mechanical analysis that the fractions of a soil are separated by making use of their differing settling-velocities. The validity of a mechanical analysis depends on the attainment of the so-called prime particle structure, in which each ultimate particle settles independently. In very fine suspensions, a change can be brought about whereby the separate particles aggregate into larger compound structures, which settle at a greater rate than that of the particles of which they are composed. This change, called coagulation, can be induced by the addition of electrolytes. For example, the addition of a few c.c. of normal calcium chloride to a litre of clay suspension is rapidly followed by the formation of floccules which, in a few minutes, settle to the bottom, leaving a clear supernatant liquid. The stability of suspensions depends on their particles being free to settle independently of each other.

The particles of a suspension carry electrical charges, usually negative, and the condition for independent behaviour is that their potential shall be above a certain critical value. When, for any reason, the potential of the particles falls below this value, the electrostatic repulsion is insufficient to prevent the association of particles into compound aggregates. The nature of the force whereby such aggregates are held together has been differently explained as cohesion, as co-ordinate valency, or as linkage through common cations.

Two stages are distinguished in coagulation. The first stage, which is reached when the potential of the suspended particles falls below the critical potential, is slow coagulation. If the potential is further lowered, coagulation becomes in-
creasingly rapid until it reaches a maximum in instantaneous coagulation, or flocculation.

Coagulation of suspensions results from a decrease in the electrokinetic potential of their suspended particles and can be brought about in one of the following ways: (1) By addition of electrolytes to the suspension medium. This results in a repression of the dissociation of the outer cations, whereby cations moving freely in the solution rejoin the colloidal particle, with the consequence that a proportion of the free charges on the inner sheath anions is neutralized. (2) By decreasing the degree of hydration of the cations, as for example, by addition of alcohol. This has the effect of decreasing the radial distance between the inner and the outer sheaths. The same effect is produced when highly hydrated cations such as lithium and sodium are replaced by less hydrated cations such as potassium, calcium, or hydrogen. (3) By increasing the dielectric constant of the medium.

The stability of suspensions in terms of their associated cations is in the order \( \text{Li} > \text{Na} > \text{K} > \text{Rb} > \text{Cs} > \text{Mg} > \text{Ca} > \text{Sr} > \text{Ba} > \text{H} \), and this is also, apart from the uncertainty in the case of hydrogen, the order of hydration of the ions.

Since the stability of suspensions depends on the charge of their particles, it follows that, so far as we are dealing with negative suspensions, the more marked is the acidoid character of the colloidal particles, the greater, ceteris paribus, will be the electrical charge and the higher its electrokinetic potential. This is, in fact, the case: clays having a high silica-sesquioxide ratio are more stable than clays having a low silica-sesquioxide ratio. Thus, whilst it is possible to prepare stable suspensions of highly siliceous clays with hydrogen as the cation, with clays of a lateritic character, i.e., with a low silica-sesquioxide ratio, the hydrogen clays are unstable, and stability can only be attained by the introduction of a highly hydrated ion such as sodium.

A review of the present position of the theory of coagulation is given by E. W. Russell.
CATAPHORESIS

One of the consequences of the electrical charge possessed by suspended colloidal particles is that they move in an electrical field, the speed of migration being proportional to the electrokinetic potential of the charged particle. Measurements of the speed and direction of cataphoretic movement, made with the aid of the ultra-microscope, have proved of the highest value in the study of the variations in the electrical properties of colloidal suspensions of differing composition.

ELECTRODIALYSIS

The exchangeable bases of soils may be separated by electrodialysis, a procedure which is essentially similar to the electrolysis of electrolytes. In the method as originally developed by S. Mattson\textsuperscript{34}, a three-compartment cell was used, the soil suspension being in an inner chamber separated from the outer compartments, containing the electrodes, by parchment membranes. The voltage used was 220 volts (D.C.) and the current was regulated by resistances in order to avoid too great a rise in temperature. The exchangeable cations appear in the cathode chamber, whilst the diffusible anions appear in the anode chamber. After 22 hours' dialysis, a Sharkey clay soil gave 28.937 mg. equivalents of base in the cathode chamber, and 3.375 mg. equivalents of acid in the anode chamber. At the conclusion of dialysis, the base-free hydrogen soil remains in the central chamber together with any indiffusible ions which may have been liberated.

Many simplified methods of electrodialysis have been proposed, for example, that of M. L. M. Salgado and G. W. Chapman\textsuperscript{35}, in which the soil is contained in a Jena glass filtering crucible through which water is allowed to percolate. The anode is a spiral of gold wire and rests on the soil, whilst the cathode is of copper gauze and is held below the filter disk. Such a method permits only the separation of cathodic electrodialysate.
The content of exchangeable bases determined by electrodialysis is generally in good agreement with that determined by displacement methods. It has been objected by some workers that electrodialysis results in a decomposition of the anionic portion of the absorbing complex. M. Trénel\textsuperscript{36} found that permutite gels, bereft of bases by electrodialysis, underwent a certain decomposition into silicic acid and sesquioxides. The desaturation thus attained is therefore not completely reversible.

AMPHOTERIC CHARACTER OF THE COLLOIDAL COMPLEX

The behaviour of the colloidal particles of the absorbing complex of soils in cationic exchanges is consistent with the theory that their outer layers consist of cations—the so-called exchangeable cations. It is not necessary to assume that the exchangeable cations are all spatially exterior. In the case of montmorillonite, with relatively wide spaces between the lattice sheets, a considerable proportion of the exchangeable cations may be attached to interior surfaces. Whether exterior or interior they will be associated with corresponding anions, which may be regarded as forming an inner layer. It is of interest to examine whether anions like cations can also exhibit exchange reactions.

The absorption of anions by soil colloids has long been established. This absorption is readily demonstrable in the case of certain anions such as $\text{PO}_4^{3-}$, but the work of S. Mattson\textsuperscript{37} has shown that even anions such as chloride which are not absorbed by ordinary soils may be absorbed by colloids having high isoelectric points.

This behaviour is illustrated by a series of experiments in which equal weights of soil colloids desaturated by electrodialysis were treated with solutions of ammonium chloride, ammonium sulphate, and ammonium phosphate, respectively, the proportion of cation to anion being varied by adding different amounts of ammonia and the respective acid.

Some typical data are given in Tables XI and XII. The data in the first table refer to a ferrallitic clay having a mole-
cular SiO$_2$/R$_2$O$_3$ ratio of 0.31, whilst those in the second table refer to a siliceous clay having a molecular SiO$_2$/R$_2$O$_3$ ratio of 3.18.

**Table XI. — Absorption of NH$_4$ and Cl Ions by Nipe Colloid, 3% Suspension (Mattson)**

<table>
<thead>
<tr>
<th>Mg. equivalents added</th>
<th>Mg. equivalents absorbed</th>
<th>Cataphoresis $\mu$/sec. at 1 volt/cm.</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>NH$_4$</td>
<td>Cl</td>
<td>NH$_4$</td>
<td>Cl</td>
</tr>
<tr>
<td>2.242</td>
<td>2.00</td>
<td>0.039</td>
<td>0.00</td>
</tr>
<tr>
<td>2.121</td>
<td>2.00</td>
<td>0.021</td>
<td>0.003</td>
</tr>
<tr>
<td>2.00</td>
<td>2.10</td>
<td>0.014</td>
<td>0.011</td>
</tr>
<tr>
<td>2.00</td>
<td>2.20</td>
<td>0.012</td>
<td>0.024</td>
</tr>
<tr>
<td>2.00</td>
<td>2.40</td>
<td>0.006</td>
<td>0.038</td>
</tr>
<tr>
<td>2.00</td>
<td>2.40</td>
<td>0.007</td>
<td>0.059</td>
</tr>
</tbody>
</table>

**Table XII. — Absorption of NH$_4$ and Cl Ions by Sharkey Colloid, 2.8% Suspension (Mattson)**

<table>
<thead>
<tr>
<th>Mg. equivalents added</th>
<th>Mg. equivalents absorbed</th>
<th>Cataphoresis $\mu$/sec. at 1 volt/cm.</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>NH$_4$</td>
<td>Cl</td>
<td>NH$_4$</td>
<td>Cl</td>
</tr>
<tr>
<td>4.178</td>
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<td>0.661</td>
<td>0.00</td>
</tr>
<tr>
<td>3.089</td>
<td>2.00</td>
<td>0.376</td>
<td>0.00</td>
</tr>
<tr>
<td>2.00</td>
<td>2.00</td>
<td>0.111</td>
<td>0.001</td>
</tr>
<tr>
<td>2.00</td>
<td>2.20</td>
<td>0.090</td>
<td>0.001</td>
</tr>
<tr>
<td>2.00</td>
<td>2.40</td>
<td>0.081</td>
<td>0.001</td>
</tr>
<tr>
<td>2.00</td>
<td>2.80</td>
<td>0.066</td>
<td>0.004</td>
</tr>
</tbody>
</table>

In both cases, with decrease in pH, cationic absorption decreases and anionic absorption increases. There is also a decrease in the electronegative character of both colloids as shown by their cataphoretic velocity; but whilst, in the case of the Nipe colloid, there is a transition through an isoelectric point to an electropositive suspension, in the case of the Sharkey colloid, although the negative cataphoresis...
decreases, no isoelectric point is indicated. Similar results were obtained for the absorption of SO$_4^-$ and PO$_4^{3-}$ ions, the energy of absorption being in the order Cl$<\text{SO}_4^->\text{PO}_4^{3-}$.

In general, colloids having low silica-sesquioxide ratios (i.e., acidoid-basoid ratios) more readily absorb anions and pass over to the electropositive state than colloids having high silica-sesquioxide ratios.

S. Mattson and A. J. Pugh$^{38}$, from an extended study of isoelectric precipitates of various hydroxides, have shown the increase in anionic absorption with iso-electric pH over a range from stannic hydroxide, isoelectric at pH 3.9, to lead hydroxide, isoelectric at pH 11.0. S. Mattson and J. B. Hester$^{39}$ found that exchange alkalinity could be developed by treating a soil with a solution whose pH had been adjusted below the isoelectric pH of the soil colloid, the anion of the salt displacing hydroxyl-ions from the colloid.

The displacement of anions from the absorbing complex has been less studied than that of cations, but recent investigations have demonstrated its occurrence. A. Demolon and E. Bastisse$^{40}$ have shown that certain anions, for example, citrate, silicate, and oxalate ions, can produce exchange alkalinity, due to displacement to hydroxyl-ions. They also demonstrated the liberation of phosphate ions and suggested that silicate ions may play a part in the mobilization of phosphate in the soil solution. S. Ravikovitch$^{41}$, in studies of the absorption and liberation of phosphate ions by soils, found that phosphate can be liberated by addition of calcium hydroxide or sodium hydroxide to soils. The liberation of phosphate ions takes place after cationic exchange has proceeded to completion and an excess of hydroxyl-ions is absorbed by the complex.

The complete analogy of anionic exchange with cationic exchange has not yet been demonstrated and, so far as soil reactions are concerned, given the electronegative character of ordinary soil colloids, appears improbable.

The electrical charge on clay particles has been discussed by R. K. Schofield$^{42}$. At a given pH value a clay will carry
a certain proportion of cations, corresponding with its total negative charge. In so far as these cations represent charges arising from isomorphous replacements in the lattice, such as the replacement of silica by tetrahedral alumina (cf. p. 131) their total amount per 100 g. of clay is constant. With varying pH, however, other groups may become active and the total charge may thus vary. Considering first the effect of increasing the pH, beginning a little below the neutral point, it is found that the increase in OH\(^-\) concentration for a given addition of alkali is less than would correspond with the amount added. According to Schofield the following reaction probably occurs:

\[ \equiv\text{Si} - \text{OH} + \text{Na}^+ + \text{OH}^- \rightarrow \equiv\text{Si} - \text{O} + \text{Na}^+ + \text{H}_2\text{O} \]

which means that a hydrogen ion dissociates from a hydroxyl, whereby the clay acquires a further negative charge and a corresponding exchangeable cation. Such hydroxyls occur at the broken edges of silica sheets and their number may be increased by breaking down the clay crystals as in fine grinding, which can be shown experimentally to increase the base exchange capacity of clays.

Schofield points out that pH in itself does not precisely fix the charge because in salt-free suspensions the exchangeable cations form a diffuse outer shell of an electrical double layer. There is thus a potential difference between the surface of the particles and the water itself and the hydrogen-ion concentration close to the particles is not the same as that found by the ordinary pH determination. This difficulty can be overcome by using suspensions in salt solutions when the thickness of the diffuse double layer is reduced to a minimum. Schofield uses normal KCl for this purpose.

Schofield found that 100 g of bentonite clay completely desaturated by treatment with dil. HCl reacts with about 100 m.e. of alkali in reaching neutrality. An equivalent amount of acidity can be extracted by leaching with strong KCl solution at pH 1.5. There appears to be a constant negative charge from about pH 6 down to pH 1.5. This he calls the permanent charge. Above pH 6, OH groups begin
to dissociate and additional negative charges are developed at higher pH values.

With certain clays there is no constant charge, because on the acid side of neutrality certain negatively charged groups lose a hydrogen ion, thereby becoming neutral, as for example:

\[-\text{Al-OH} \rightarrow \text{Al=O} + \text{H}^+\]

When this occurs the clay can absorb chloride from alcoholic HCl, the amount of chloride absorbed increasing with decrease in pH to a maximum which, however, may not be sufficient to render the particles electropositive.

If the basic groups exceed the isomorphous replacements responsible for the permanent negative charge, then below a certain pH the positive charges would exceed the permanent negative charge. At this pH, the isoelectric point, the clay is electrically neutral and no absorption is possible.

It is well known that clays of the montmorillonite-beidellite type have higher base exchange capacity than clays of the kaolinitic type. In the latter type, isomorphous replacement of silica by tetrahedral alumina scarcely occurs and most of the base exchange is attributable to hydroxyls attached to silicones at the edge of the silica layers. The increase of base exchange capacity by fine grinding is thus explained.

The ultimate pH of a soil, i.e., the pH after all diffusible anions and cations have been removed, as in electrodialysis, depends on the relative dissociation of the acidoid and basoid components. Where \( K_{\text{acidoid}} > K_{\text{basoid}} \), the reaction is on the acid side of neutrality. This is the case even with the most markedly basic soil colloids such as that of the Nipe clay. Mattson (loc. cit.) obtained the following pH values for suspensions of four typical colloids after electrodialysis.

<table>
<thead>
<tr>
<th>Clay</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nipe</td>
<td>6.65</td>
</tr>
<tr>
<td>Cecil</td>
<td>4.75</td>
</tr>
<tr>
<td>Sassafras</td>
<td>4.63</td>
</tr>
<tr>
<td>Sharkey</td>
<td>3.73</td>
</tr>
<tr>
<td>Humus</td>
<td>3.72</td>
</tr>
</tbody>
</table>

The corresponding SiO\(_2\)/R\(_2\)O\(_3\) ratios for the four
mineral colloids were 0.31, 1.34, 1.89, and 3.18, respectively. M. S. Anderson and H. G. Byers found rather lower values for the pH of electrodialyzed colloids, namely, 2.5-3 for siliceous and 3.5-4.5 for sesquioxidic colloids.

We have seen that Mattson considers the soil colloids to be analogous to isoelectric precipitates. The colloidal material of the soil, according to this hypothesis, has a complex constitution and varies in properties according to the proportion of acidoid to basoid groups. The principal acidoid groups are silicic acid and humic acids, and the principal basoid groups, alumina and ferric oxide. We may contrast colloids of an acidoid character having low isoelectric pH, low ultimate pH, feeble anionic absorption, and high base exchange capacity, with colloids of basoid character, having high isoelectric pH, high ultimate pH, strong anionic absorption, and low base exchange capacity. The former are represented by colloids rich in silicic acid and humus, the latter by colloids rich in sesquioxides.

The problem of the constitution and reactions of the soil colloidal material is being actively investigated from different standpoints and a full discussion would lead into regions of physical chemistry beyond the scope of the present book. Mattson and his school have attacked the problem from the standpoint of the theory of isoelectric precipitates and amphoteric compounds, starting from the classical studies of Loeb and Michaelis. A. J. Pugh has attempted to link up this line of enquiry with the electronic theory of valency. (Cf. p. 168.)

On the other hand, there is the approach to the problem through the study of the crystal structure of the clay minerals. For example, L. Wolf and R. Kächele have explained the cationic exchange phenomena of acid-treated kaolin and ceramic clays as connected with the displacement of aluminium from exposed parts of the crystal lattice. C. E. Marshall distinguishes two classes of exchangeable cations, namely, those outside the clay particle forming part of the electrical double layer and those in the spaces between the
lattice sheets, the latter predominating in the beidellite-montmorillonite group of minerals. Marshall agrees with Wolf, whose experiments relate to kaolins, in regarding what was formerly thought to be a hydrogen clay as an aluminium clay.

It appears from experiments by R. Chaminade and G. Drouineau\(^{47}\) that reactions can take place whereby certain cations can pass into the non-exchangeable state. Such "reverted" cations can be brought into the exchangeable state by grinding. This behaviour is shown by magnesium and potassium but not by calcium ions. The exchange property of clay may be located in hydroxyl or substituted hydroxyls. W. P. Kelley, H. Jenny, and S. M. Brown\(^{48}\) would locate it in sorbed water molecules falling into two categories, namely, (1) broken-bond water at fractured surfaces and (2) planar water held by stray electric fields at the surfaces of Si–O–Si or OH–Al–OH planes. The latter water is more loosely held and is given off below 150°C. It forms the greater part of the sorbed water of 2:1 minerals such as montmorillonite. Broken bond water predominates in 1:1 minerals and in felspars, which develop base exchange properties in grinding. There is a linear relationship between water lost at 150°C and base exchange capacity for a range of soil colloids.

G. Wiegner\(^{49}\) has emphasized the importance of metastructure, the actual form of the broken surfaces, in determining the magnitude of exchange capacity.

The crystal concept of the absorbing complex and the isoelectric precipitate theory appear to be in conflict. Mattson, whilst admitting the crystalline character of clay, postulates an indefiniteness of composition in the active surface layers. The possible replacements suggested by Marshall (p. 131) imply a criticism of the supposed correlation between SiO\(_2\)/R\(_2\)O\(_3\) ratio and soil properties. Yet abundant experimental data have shown the connexion between this ratio and soil behaviour. Is it possible to consider sesquioxidic colloids as consisting of clay minerals of the montmorillonite-beidellite type with hydrated sesquioxides
as diluents or do they contain characteristic minerals of a less siliceous character than those of the beidellite-montmorillonite group? Until extended crystallographic studies have been carried out on soils of primary weathering, including the highly sesquioxidic soils of the tropics and sub-tropics, it will be impossible to give a final verdict between the rival theories.

PHOSPHORUS COMPOUNDS IN SOIL

Soils contain varying proportions of phosphorus, present in a number of different categories, inorganic and organic. Ultimately all the phosphorus in soils originates from the apatite minerals of igneous rocks. The general formula of apatite is \( \text{Ca}_10(\text{PO}_4)_6\text{X}_2 \), where \( \text{X}_2 \) represents \( \text{F}_2, \text{Cl}_2, \text{CO}_3 \), or \( \text{(OH)}_2 \). Some of the apatite present may be primary apatite surviving from an ancestral igneous rock, but in most soils apatite, if present, is of secondary origin.

Up to the present no completely satisfactory fractionation of soil phosphorus has been devised, but the following categories may be present:

1. Apatite minerals.
2. Dicalcium phosphate, \( \text{CaHPO}_4 \), generally a fugitive constituent originating from the reaction of soluble phosphates with calcium carbonate. It tends to revert to apatite or, in acid soils, to ferric and aluminium phosphates.
3. Ferric and aluminium phosphates. These are the most stable inorganic phosphates in acid soils. In neutral or alkaline soils they tend to pass over to apatite.
4. Phosphoric acid groups in the colloidal complex. These may originate during the process of weathering or the biological decomposition of organic phosphorus compounds. They may also enter the complex from soluble phosphates added in fertilizers.
5. Phosphoric acid groups in humus. R. Chaminade has demonstrated the occurrence in normal fertile soils of absorption compounds of phosphoric acid and humus. Below pH 5.5 they tend to dissociate and are therefore absent from
acid soils. Phosphoric acid present in this form in fertile soils is fairly stable and does not readily undergo reversion to insoluble and unavailable forms. The existence of the compounds would appear to depend on the presence, associated with the humus, of calcium ions through which the absorption may take place.

6. Organic phosphorus compounds such as nuclein and lecithin present in undecomposed plant residues or in manures of animal origin, such as meat guano.

The phosphorus pattern of soils is strongly affected by their reaction. In acid soils the tendency is for aluminium and ferric phosphates to be the principal mineral phosphates whilst in neutral and alkaline soils, apatite tends to be formed. G. Nagelschmidt and H. L. Nixon have examined fragments of chalk from plots at Rothamsted that had received manurial dressings including superphosphate for over a century. After ignition at 800\(^\circ\) and removal of calcium oxide by extraction with sucrose solution, a residue was left, containing 47\% CaO, 26\% P\(_2\)O\(_5\), and 1.5\% F. It gave the X-ray powder diagram of apatite. R. Williams found the greater part of the total phosphorus of acid soils to be soluble in sodium hydroxide solution. This was considered to consist of the phosphorus of iron and aluminium phosphates together with organic phosphorus. The small proportion insoluble in sodium hydroxide probably represents apatite phosphorus.

The fixation of phosphate ions by soils, first demonstrated by Way, has long been studied. For example, E. J. Russell and J. A. Prescott showed that the partition of phosphate ions between soil and solution in the presence of dilute acids obeys the ordinary absorption equation \(y = KC^1\). The analogy with base exchange, however, is not complete since soluble phosphate in the presence of soil can be removed by ordinary precipitation. Insoluble phosphates can be formed by reaction with (a) calcium carbonate, (b) hydrated sesquioxides and (c) exchangeable calcium liberated by base exchange. It is only within recent years that the
possibility of phosphorus entering into the absorbing complex as an acidoid constituent analogous to silicic acid and humic acid has been envisaged.

The fixation of soluble phosphate by soils is strongly influenced by the reaction of the medium. Thus H. O. Askew found maximum fixation at pH 3.3 to 5.5 depending on soil type. A. F. Heck distinguishes three types of fixation, namely (1) readily available phosphate as $\text{Ca}_3(\text{PO}_4)_2$; (2) moderately available phosphate as $\text{AlPO}_4$; and (3) difficultly available phosphate as $\text{FePO}_4$ and $\text{Al}_2(\text{OH})_3\text{PO}_4$. Minimum fixation as "difficultly available" phosphate occurred at pH 6.3 for a siliceous type of soil and at pH 7-7.5 for a lateritic soil.

T. Gaarder, in a study of water-soluble phosphate, found the ratio of basoids ($\text{R}_2\text{O}_5$) to acidoids (silicic acid and humic acids) extractable by Tamm's acid oxalate method to have a marked effect. With an excess of basoids over acidoids there was no water solubility over a range of pH 3-8.5. With acidoids in excess, water solubility is shown over the whole range, a result in harmony with the known depletion of this constituent in the A horizons and its accumulation in the B horizons of podsol soils. The fixation of phosphate by sesquioxidic soils has been repeatedly shown by workers in many countries. A review of the work on phosphate fixation is given by P. L. Hibbard.

E. G. Williams in a discussion of phosphate fixation in soils distinguishes two types of fixation, namely biological and chemical.

Biological fixation occurs when certain micro-organisms assimilate phosphoric acid into organic form. On the death of the organisms the organic matter may undergo decomposition, with release of phosphate ions.

Chemical fixation takes place in two main ways, namely (1) by absorption, and (2) by precipitation. Absorption is the reaction of phosphate ions with certain clay minerals. The phosphate thus fixed is relatively unavailable to plants. This type of fixation is most marked in acid soils.
Chemical precipitation of phosphate is governed by soil reaction. In acid soils, phosphate is precipitated mainly as iron and aluminium phosphates, which are relatively unavailable to plants. In slightly acid to neutral soils phosphate is precipitated as calcium phosphates that are fairly available for crop nutrition. Such precipitated phosphate may correspond with the phosphate-humate complexes of Chaminade. In alkaline soils the tendency is for less available forms of calcium phosphate, possibly apatite-like minerals, to be formed.

The fixation of phosphate by soils is one of the principal problems in the use of fertilizers, since it limits the effect of added phosphate to those layers to which fertilizers have been applied and may even result in complete unavailability to plants. The most urgent problem in connexion with the phosphorus of soils is to obtain some satisfactory method of fractionation.

MANGANESE AND IRON CYCLES.

The proportion of manganese in the earth’s crust is not more than about 0.1%, mainly in the bivalent form. Manganese can occur in soils and their parent materials in bivalent, tervalent, and quadrivalent form. The oxidation-reduction equilibria involved are of particular interest in connexion with the problem of manganese deficiency in crop growth.

The bivalent manganese of soils, apart from that present in unweathered minerals, occurs as exchangeable manganese and can vary in amount from over 10 mg.% down to a mere trace. Tervalent and quadrivalent manganese mainly occur as oxides or their hydrates. Manganese dioxide is a familiar constituent in soils subject to alternate oxidizing and reducing conditions and may occur in a variety of forms from small black flecks to massive concretions.

The forms and reactions of manganese in the soil have been discussed by G. W. Leeper. According to this writer the following processes are concerned in the manganese
cycle in soils: (a) weathering of minerals and decomposition of plant residues to produce bivalent manganese; (b) downward movement of manganous ions in drainage water; (c) equilibria between manganous ions in the soil solution and those in the soil colloids; (d) absorption of manganous ions by plants and return of the same to the soil in residues; (e) oxidation, mainly biological, of manganous ions to give higher oxides; (f) ageing of active higher oxides to relatively inert forms; (g) reductions of higher oxides to manganous ions by organic matter or by anaerobic bacteria, and (h) direct absorption of manganic ions by plants or micro-organisms.

Under acid conditions the higher oxides of manganese are reduced to manganous forms. This reduction is mainly biological and is favoured by the presence of oxidizable organic matter and by anaerobic conditions. Under alkaline conditions the tendency is for the higher oxides of manganese to be formed.

According to H. G. Dion and P. J. G. Mann, a significant part of soil manganese may be in the form of hydrated manganic oxide formed by the oxidation of manganous forms. This can undergo dismutation into manganese dioxide and bivalent manganese under weakly acid conditions. Manganous manganese in exchangeable form can originate also by reduction of manganese dioxide. The amount of each form present will be the result of an equilibrium depending on the relative speeds of oxidizing processes on the one hand and reducing and dismuting processes on the other, the latter being less active at higher pH values.

Iron compounds in the soil readily undergo oxidation and reduction. The grey and bluish-grey colours of soils under continuous anaerobic conditions indicate the presence of ferrous compounds such as ferrous sulphide, FeS₂, vivianite, Fe₃(PO₄)₂, etc. Ferrous hydrogen carbonate formed under reducing conditions is relatively soluble and may be removed by drainage, generally in a lateral direction.
owing to restriction of downward percolation. This leads to an impoverishment of the soil in iron and is partly responsible for the grey colours of wet soils. The reduction of ferric iron to ferrous is favoured by the presence of organic matter and, to some extent, by the neutral or alkaline reaction. The process, as viewed in the field, is sometimes termed gleysing. Where reducing and oxidizing conditions alternate as in the zone of a fluctuating water-table ferrous compounds formed by reduction may be again oxidized to give hydrated ferric oxides, which appear as streaks, mottlings, or, in extreme cases, as massive precipitates, as in bog iron ore. Sulphuric acid from the oxidation of ferrous sulphide or iron pyrites may give exceedingly acid soil reactions (cf. p. 159).

**OXIDATION-REDUCTION POTENTIAL OF SOILS**

Many of the reactions of soils are connected with the relative intensity of oxidizing and reducing conditions within the soil. According to the electronic theory, in oxidation, an atom yields up one or more electrons and thereby becomes positively charged. Conversely, in reduction, an atom receives one or more electrons and becomes negatively charged, e.g., Fe²⁺→e⁻→Fe³⁺. It is possible to arrange oxidizing and reducing agents in the order of their **oxidation-reduction potential** by forming an electrical cell with each system and an arbitrary system, e.g., a calomel cell, and measuring the potential set up between them. Numerous attempts have been made to determine the oxidation-reduction potential of soils. The results obtained are strongly influenced by the methods used and it cannot be said that any single method has established itself. The results hitherto obtained do not appear to be of great significance. A discussion and comparison of methods is given by L. A. Brown.⁶¹

S. C. Heintze⁶² has demonstrated the relationship between pH and Eh (oxidation reduction potential) by curves. It is considered improbable that reversible oxidation-reduction equilibria are established in normal soils or that Eh values can be used for anything more than diagnosing
acutely reducing conditions. There is no poising system analogous to the exchange complex that buffers acid-base reactions in soils. A. N. Puri and A. Sarup\(^{83}\) consider that the relationship between pH and Eh is so close that pH suffices to indicate Eh.

**BIBLIOGRAPHY**


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CHAPTER VII

SOIL ORGANIC MATTER

ORIGIN FROM PLANT RESIDUES

In our general view of the constitution of the soil, we have discussed the relationship of organic matter to the other constituents present. We must now consider its constitution in connexion with those changes effected by micro-biological agencies, whereby plant residues become incorporated with the soil as humus.

The proportion of organic matter in the soil varies considerably. Whilst in the case of peat soils, the dry matter may be almost entirely organic, in ordinary cultivated soils, organic matter rarely amounts to more than 10% of the total dry weight. The highest figures, apart from those in peats, are obtained in grassland soils under humid climates, and the lowest figures under deficient rainfall in hot climates.

The variations in organic matter content in the soil profile are also considerable. There is a general decrease in organic matter from the surface downwards in all soils, except humus podsols with horizons of humus accumulation. Whilst the decrease in organic matter with depth is generally gradual, in some cases there is an abrupt decrease at a certain point in the profile, notably in soils with impeded drainage, where there is often a sharp transition from the humic top soil to the raw subsoil.

The relationship of the organic matter content of the soil to such factors as climate, soil texture, water conditions, and cultivation will be more appropriately discussed when the mechanism of the decomposition of plant residues has been examined.
The humus of soils arises mainly from plant residues. There is, of course, a certain contribution from animal remains and excreta, notably in the case of cultivated soils. Under forest vegetation, the principal source of the soil organic matter is leaf-fall. In certain types of forest, however, undergrowth and ground vegetation make a considerable contribution from their residues. Indeed, it is probably only in the case of artificial coniferous plantations with close canopy that leaf-fall forms the sole source of organic matter. Beech forests which, during the summer months, may be devoid of ground vegetation, may yet have an appreciable ground flora in spring.

Under steppe or prairie conditions, the soil is occupied by graminoeous and herbaceous plants, whose growing season is limited by drought or winter cold. The dead remains of each season's growth become added to the soil in the form of leaves, stems, and roots. Artificial grassland, in this respect, can be classed with steppe and prairie. Under cultivation, organic matter is added to the soil in the form of stubble, trash, and dressings such as farmyard or pen manure.

Plant residues added to the soil differ in certain respects from fresh plant materials. They represent, in the main, structures which have fulfilled their functions and are generally almost devoid of starches, sugars, and fats; for these have already been stored in seeds or tubers, or have been transformed into fibrous materials. The principal constituents of plant residues, then, are old structural elements, together with small amounts of protein, traces of waxes and resins, and ash constituents.

Oat straw, which is typical of all plant residues, contains, according to H. Pringsheim¹:

<table>
<thead>
<tr>
<th>Component</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Celluloses</td>
<td>35.43</td>
</tr>
<tr>
<td>Pentosans</td>
<td>21.33</td>
</tr>
<tr>
<td>Lignins</td>
<td>20.40</td>
</tr>
<tr>
<td>Crude Protein</td>
<td>4.70</td>
</tr>
<tr>
<td>Waxes</td>
<td>2.02</td>
</tr>
<tr>
<td>Ash</td>
<td>4.81</td>
</tr>
</tbody>
</table>

The elemental composition of some plant residues is shown by the following data given by Dvůrák². They relate to dry matter in each case.
SOILS

<table>
<thead>
<tr>
<th></th>
<th>C</th>
<th>H</th>
<th>N</th>
<th>O</th>
<th>Ash</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pine needles</td>
<td>41.96</td>
<td>3.98</td>
<td>1.42</td>
<td>21.07</td>
<td>31.57</td>
</tr>
<tr>
<td>Oak leaves</td>
<td>49.11</td>
<td>6.12</td>
<td>1.71</td>
<td>29.38</td>
<td>13.68</td>
</tr>
<tr>
<td>Wheat Straw</td>
<td>47.01</td>
<td>5.66</td>
<td>0.82</td>
<td>38.61</td>
<td>7.90</td>
</tr>
<tr>
<td>Lucerne</td>
<td>43.28</td>
<td>5.86</td>
<td>1.95</td>
<td>38.54</td>
<td>10.37</td>
</tr>
</tbody>
</table>

MICRO-ORGANISMS OF THE SOIL

Whilst purely chemical decompositions play a certain part in the decomposition of plant residues, particularly in tropical soils, these changes are mainly due to the activity of the soil micro-fauna and micro-flora. The more important groups of the soil micro-fauna are:

I. Protozoa.
II. Nematodes.
III. Worms, insects, etc.

The more important groups of the soil micro-flora are:

I. Algae and Diatoms.
II. Fungi, including Actinomycetes.
III. Bacteria.

A. Autotrophic, deriving their energy from the oxidation of simple inorganic compounds and their carbon from carbon dioxide. They include:
   i. Nitrifying bacteria.
   ii. Sulphofying bacteria.
   iii. Iron bacteria.

B. Heterotrophic, deriving their carbon from complex organic compounds. They include:
   i. Nitrogen fixing bacteria.
      a. Symbiotic, e.g., Bact. radicipila.
      b. Free living.
         1. Aerobic, e.g., Azotobacter.
         2. Anaerobic, e.g., Clostridium.
   ii. Bacteria concerned with the process of ammonification.
   iii. Cellulose bacteria and other bacteria decomposing fibrous materials.
The numbers of bacteria in soils have been estimated by methods which it is not necessary to describe here. That the figures obtained represent actual conditions in the natural soil may be received with some doubt; but they may serve as a basis for comparison. In ordinary soils from 2 to 200 millions of bacteria per gram of soil have been recorded, the highest figures being obtained in highly cultivated and heavily manured soils.

The distribution of bacteria within the soil itself is essentially sporadic. C. Thom distinguishes between the explosive decomposition associated with the presence of fresh material and the slower decomposition in the soil as a whole associated with the characteristic soil organic matter or humus. The intense activity in the former case is associated with enormous increases in bacterial numbers. There is a marked localization of bacterial activity in the vicinity of growing roots, where epidermal and cortical cells, as well as dead root hairs, are continually being sloughed off.

Considerable seasonal fluctuations are observed in the bacterial numbers in soils. The highest figures are obtained under the joint effect of moisture and warmth, whilst drought and cold, particularly the latter, exert a depressing effect. There is a rapid decrease in bacterial numbers with depth, and complete sterility is reached within a few feet of the surface.

According to C. B. Taylor the general character of the autochthonous micro-flora is relatively uniform in a soil of a definite type, even although productivity may be altered by manuring. So far as the soil micro-population plays a part in pedogenesis and in the metabolism it should be regarded not as an independent factor, but as itself conditioned by the physical and chemical properties of the soil.

Fungi are present in much smaller numbers than bacteria. They are most abundant in organic soils with acid reaction. In such soils the decompositions are mainly the work of fungi rather than of bacteria.
The work of the lower animals may be considered in the first place. Under British conditions, this type of attack is mainly through the agency of earth-worms, who feed on the soil, and, by passing it through their bodies, perform an important task in comminuting and mixing the plant residues with the mineral portion of the soil. It has been calculated by Darwin that, in the course of half a century, the whole of the soil of a pasture down to ten inches has been brought to the surface in the form of casts. It is generally considered that the rôle of earth-worms in the decomposition of organic matter in the soil is mainly mechanical.

H. Franz considers that all the smaller soil fauna play a part in the transformations of soil organic matter. Nematodes are particularly numerous, often amounting to about 200 per c.c., and are active in the decomposition of cellulosics. Whilst earthworms are the most active agents, such organisms as collembola larvae and caterpillars also play a part.

In soils with markedly acid reaction, earth-worms are present in smaller numbers, and the residues of plants tend to accumulate at the surface as a raw humus or duff layer. This is seen in the podsols of coniferous forests and heaths.

In the tropics, the most important invertebrate connected with the organic matter cycle is the termite or white ant.

The changes induced by earth-worms, white ants, and similar organisms, although they involve certain chemical decompositions, are principally of importance through their effect in distributing organic matter throughout the soil profile. In the absence of such organisms, as in highly acid soils, the added organic matter remains as a sharply defined layer resting on a mineral soil which is generally deficient in organic matter.

By reason of their enormous numbers, it is the organisms constituting the micro-flora which are of chief importance in the decompositions affecting soil organic matter. Before proceeding to an account of their activities it may
be helpful to draw certain distinctions between their conditions of development and function.

The first distinction to be drawn is between autotrophic and heterotrophic organisms. Autotrophic organisms are those which obtain their carbon from carbon dioxide and their energy from the oxidation of inorganic compounds. Examples of these are the nitrifying bacteria.

Heterotrophic organisms, on the other hand, obtain their carbon from complex organic compounds and their energy from the oxidation of such compounds. Included in these are the bacteria which oxidize cellulosic materials, and the bacteria concerned in the fixation of atmospheric nitrogen. Heterotrophic bacteria may be divided into aerobic organisms working in the presence, and anaerobic organisms, working in the absence or deficiency of oxygen.

The transformations affecting soil carbon and soil nitrogen are the work of autotrophic and heterotrophic, of aerobic and anaerobic organisms. Grossly considered, these changes may be grouped into two classes, namely, oxidative decompositions and humification.

OXIDATIVE DECOMPOSITION OF ORGANIC MATTER

In oxidative decomposition, organic matter is oxidized to carbon dioxide and water. These changes, which are the work of many organisms, are favoured by good aeration, and are most rapid between 35 and 40°C. Moisture is necessary, since bacteria live in a fluid medium, but excess of moisture by excluding air from the soil, slows down and may even inhibit bacterial oxidation. The presence of calcium carbonate would appear to favour oxidative decomposition of organic matter; but it is by no means certain that oxidative decomposition cannot proceed equally well in the absence of calcium carbonate, provided there is a sufficient supply of calcium, combined in readily available forms with the clay and organic matter of the soil, to serve as a source of calcium and to maintain a reaction favourable to bacterial life. A high calcium-status, by favouring granular
structure, improves aeration. Soils containing large amounts of free calcium carbonate are particularly well aerated and oxidation of organic matter proceeds in them with great rapidity, provided adequate moisture be present.

The oxidative decompositions will therefore proceed with the greatest rapidity where optimum moisture conditions, optimum temperature, and a sufficiency of lime in the soil coincide. Under such conditions, plant residues added to the soil readily disappear, a circumstance which is reflected in the low organic matter content of cultivated soils of hot countries. Even in colder climates, the aeration induced by cultivation can lead to the rapid disappearance of organic matter from the soil. An example of this is given by some data obtained by the writer in North Wales.

A cultivated soil from the uplands of Caernarvonshire showed a loss on ignition of 16.6%. On the other side of the fence, the soil, which presumably corresponded with that from which the field had been reclaimed 50 years before, gave an ignition loss of 43.1%. Although ignition loss is an untrustworthy measure of organic matter, the difference between these figures does give an approximate idea of the rapid loss which has taken place, and this in spite of the fact that the cultivated soil had been from time to time under long grass leys and had received the ordinary dressings of farmyard manure during the arable break.

The oxidative decomposition of organic matter results in the liberation in the soil of large amounts of carbon dioxide. In so far as micro-biological activity can be correlated with general fertility, the rate of evolution of carbon dioxide may afford important information as to biological conditions in the soil. E. J. Russell found a correlation between the rate of absorption of oxygen and the productivity of soils in the same locality.

The carbon dioxide produced in the soil by the oxidation of plant residues is the most important single source of this compound in nature. Its significance both for the nutrition of plants and also for the processes of soil formation can
hardly be over-estimated. Dissolved in the soil moisture or combined in the form of hydrogen-carbonates, it plays an important part in the vertical translocation of materials within the soil profile. In the absence of humic acids from peaty organic matter, carbonic acid is the principal acid concerned in soil formation.

Whilst the oxidative decomposition of organic matter in the soil is mainly the work of micro-organisms, it is probable that purely chemical decompositions also play a part. Such decompositions are likely to be of greatest moment under tropical conditions in soils subjected to periods of such intense drought as to suppress micro-biological activity. Under humid temperate conditions their part is probably subsidiary to the decompositions effected by micro-organisms.

It should be noted that oxidative decomposition and, indeed, all micro-biological decompositions are more rapid in the case of plant residues than in the case of the actual humus, which is relatively stable and resistant to micro-biological and chemical oxidation.

HUMIFICATION

We have seen under what conditions added organic matter becomes destructively oxidized. There is another class of processes whereby, although some destructive decomposition takes place, a considerable proportion of the added organic matter survives in the form of a dark-coloured amorphous material, which may be termed humified organic matter or humus. There are at least four distinct types of humification, which will be examined in turn.

1. Anaerobic Humification. This type of decomposition is shown most typically in the changes undergone by the residues of aquatic vegetation in ponds and lakes. The gaseous products are methane, hydrogen, and carbon dioxide, and there is at the same time a production of amorphous dark material which, by continued accumulation,
forms fen or lacustrine peat. In ordinary soils, this type of decomposition is intermittent. It predominates when soil aeration is defective, as during the winter months, or indeed at any time when the soil is very wet. It may be that, even when the general conditions in the soil favour aerobic decomposition, the conditions within the actual soil crumbs favour anaerobic decomposition. The humus of normal soils under humid temperate conditions must originate mainly from anaerobic decompositions.

The effect of temperature on anaerobic humification is somewhat masked by the fact that the higher temperatures of the summer months are generally associated with drier soil conditions, so that humification becomes recessive. But when, owing to subsoil water conditions or to heavy summer rainfall, the soil is kept in a permanently waterlogged condition, it may be presumed that the humification processes will be, equally with the oxidative processes, favoured by warmth. The development of peaty organic matter occurs, even in the tropics, in waterlogged situations.

Anaerobic humification goes on both in calcareous and non-calcareous soils under permanent waterlogging. When, however, waterlogging is only seasonal, the presence of calcium carbonate by promoting an open texture and free drainage, will tend to abridge the period of anaerobic decomposition.

The extreme result of anaerobic humification is seen in bog peat, the condition for whose formation is the prevalence of permanent wetness, either from the presence of a high water-table, or from surface seepage, as in high-moor and mountain peats. The origin and classification of peats will be examined in greater detail in a later chapter.

II. Acid Humification. This term may be applied to those decompositions affecting organic residues in the absence or deficiency of calcium and other bases. This is exemplified by the production of raw humus under heath conditions. That humification in this case is not conditioned by anaerobic processes may be shown by the fact that, on
acid rocks such as quartzite, raw humus develops, even under conditions where aeration is most thorough. Indeed, on such rocks it may be observed that the residues of crevice vegetation, such as lichens and mosses, readily form a black residue. This type of decomposition prevails under heath and natural coniferous forest. Conifers and heath plants make comparatively small demands on the mineral reserves of the soil and their residues are correspondingly poor in basic constituents. The decomposition of these residues is to a large extent carried out by fungi. The acid conditions limit the activity of earth-worms, and the raw humus or duff accumulates as a sharply defined layer at the surface.

In the literature of forestry, it is frequently stated that certain plants are associated with the production of raw humus. The nature of this association is not clearly understood and is sometimes apprehended wrongly. Heath, Erica cinerea, and whortleberry, Vaccinium myrtillus, for example, are frequently mentioned as raw humus formers. Their association with the formation of raw humus may be attributed primarily to the circumstance that they make very small demands on the soil for mineral nutrients, and therefore tend to occupy soils which are naturally poor in such constituents, or which by degradation and impoverishment have fallen to a low mineral status. The low base-status of their residues largely contributes to the development of raw humus, the acid leachings from which still further deplete the scanty base reserves of the soils on which they are found.

Whilst it is impossible to decide whether the invasion of an acidophilous vegetation has caused soil deterioration or whether these plants have occupied a soil which has become unfit to support the growth of plants with more exacting mineral requirements, it is certain that the production of raw humus from the residues of such plants intensifies the leaching of mineral matter from soils, and still further accentuates the conditions antecedent to this formation.
The direct effect of acidity on the decomposition of organic matter in the soil is shown by some experiments of J. W. White, F. J. Holben and C. D. Jeffries in which the rate of evolution of carbon dioxide from a soil adjusted to different pH values was measured. The rate decreased with the pH. The same effect was observed when various organic materials were added to the soil, an exception being starch, for which no depression of the rate of decomposition by acidity was observed.

The raw humus formed under conditions of acid humification differs considerably from the humus formed by anaerobic processes in normal soils and that in peat soils. Raw humus is formed mainly through the activity of fungi. The soil micro-fauna plays a minor part and bacterial decompositions are also recessive. The bog peat formed under conditions of low base status, however, may possibly have some resemblance to raw humus.

III. Humification Under Semi-Arid Conditions. The increasing aeration of the soil accompanying a decrease in rainfall in passing from a humid to a drier climate is reflected in a progressive decrease in the proportion of organic matter in the soil under forest vegetation.

Where the climate is no longer sufficiently humid to carry a natural forest vegetation, and the prairie or steppe region is entered, there is a marked increase in the organic matter content of the soil. This can be verified in any locality where, under approximately the same rainfall, both steppe and forest vegetation occur. Moreover, in the steppe region, the highest contents of organic matter occur under more arid conditions than those of the zone of transition from steppe to forest.

It seems clear that the accumulation of organic matter in such soils, which include the black earths of Russia and the prairie soils of North America, must be the result of a type of humification which is essentially different from that which occurs in more humid climates. On the one hand, the scanty rainfall precludes the prevalence of anaerobic con-
ditions, except in depressions, whilst on the other hand, the high base-status prevents an acid type of humification.

Whilst the nature of the decompositions which occur under such conditions has not yet been completely investigated, it seems probable that the decisive factor is the occurrence of summer drought, resulting in a lowering of the soil moisture content below the point at which micro-biological actions can proceed.

The high base-status of the soils of the semi-arid regions also exerts an effect. It is noteworthy that, in the tropics, soils containing free calcium carbonate or a high content of exchangeable bases are usually dark in colour. It appears that a high lime-status coupled with the occurrence of high temperatures results in a darker type of humus than that produced under more humid conditions. These conditions are reproduced in soils such as the prairie soils and the continental rendzinas which, although they do not occur in definitely arid climates, possess a high base-status and are subjected to dry hot summers. The dark colour of prairie and steppe soils may be to some extent due to the presence of elementary carbon formed by the charring of organic matter through fires.

Owing to the fact that the type of humus formed under conditions of base-saturation is of a darker colour than the humus of base-unsaturated soils, a visual estimate of the organic matter content of soils of the rendzina and black earth groups may frequently lead to an exaggerated idea of the amount actually present. For example, in Barbados, S. J. Saint reports that red soils have a higher organic matter content than black soils derived from the same limestone material.

IV. Humification under humid tropical conditions. We may, perhaps, distinguish a fourth type of humification, namely, that obtaining under tropical-humid conditions. In spite of the high temperature, favourable to the oxidation of plant residues, the organic matter status of these soils is not as low as might be expected from theoretical considerations.
and also from their appearance. Little is known about tropical humus, but it may be conjectured that it is much less dark in colour than that of temperate regions. In tropical red soils some of the humus may be in the form of iron-humus complexes.

FACTORS GOVERNING THE ORGANIC MATTER CONTENT OF SOILS

The amount of organic matter in a given soil represents the balance between the addition of material from plant residues and the destructive decomposition by soil microorganisms. A low content of organic matter may result either from the scanty addition of plant residues, as in the case of desert soils, or from the rapid oxidative decomposition of organic material under the joint influence of high temperatures, free aeration, and a sufficiency of moisture for microbiological processes.

Broadly speaking, we may consider the humification processes as conservative of organic matter. The most rapid accumulation of organic matter in the soil will occur under perpetually waterlogged conditions, when the decomposition is mainly anaerobic.

High rainfall, by encouraging plant growth and maintaining soils in wet, and therefore inadequately aerated, conditions, favours the accumulation of organic matter. The relation between rainfall and organic matter content is, however, complicated by the effect of natural vegetation. Steppe or savannah is generally more favourable to the accumulation of organic matter than forest. Under forest conditions, the added residues consist principally of leaf-fall and this, as litter, is exposed to considerable losses by aerobic decomposition. Under steppe the residues are more thoroughly incorporated with the soil, and further micro-biological decomposition is frequently inhibited through lack of moisture.

The inhibition of micro-biological decomposition during the dry season may account for the considerable proportions of organic matter found in certain virgin tropical soils. At
present, little is known of the organic matter cycle under such conditions; but it is certain that cultivation and irrigation in hot climates lead to the rapid disappearance of organic matter from soils. Whilst it is not unusual to find as much as 10% of organic matter in virgin soils in the humid tropics, in irrigated soils of semi-arid regions, such as the Sudan, less than 1% may be present. Sugar cane soils in Trinidad under long cultivation may contain only 2-3% of organic matter in spite of heavy dressings of pen-manure.

The relationships of organic matter to climate and cultivation in Great Britain are fairly simple and intelligible. Grassland soils are, as may be expected, richer in organic matter than arable soils, not only because they receive a greater contribution of organic matter from the residues of vegetation and the excreta of grazing animals, but also because the intense aeration produced by tillage operations favours oxidative decomposition by aerobic organisms.

Texture also has an influence on the organic matter content of soils. Light soils are generally better aerated and warmer, and the decomposition of added organic matter proceeds more rapidly than in close-textured soils.

Apart from texture, the water-air conditions are markedly influenced by situation and, in particular, by the position of the water-table. With a high water-table, even an open-textured light soil may be maintained in a state of deficient aeration, with the resultant tendency for organic matter to accumulate as the products of anaerobic humification. Such conditions lead in extreme cases to peat formation.

The effect of calcium carbonate on the organic matter content under British conditions is somewhat complex. In the first place, the presence of free calcium carbonate confers a flocculated or crumb structure on heavy clay soils, and to that extent favours aeration and decomposition. Where calcium carbonate becomes a dominant constituent, as in chalk soils, the texture is very open and the excessive aeration is further accentuated by the tendency to shallow-
ness. It is not surprising, therefore, that calcareous soils and, in the highest degree, chalk soils tend to be deficient in organic matter. It is also possible that a high lime-status in itself may be favourable to aerobic decompositions, apart from its indirect effect through soil texture.

Variations in climate have an important effect on the balance between gains and losses of organic matter. Under high rainfalls, as in the West of England and Wales, the soil is maintained in a wet condition through a long period of the year and, whilst vegetative growth is favoured, oxidative decomposition is considerably repressed. Further, the cool summers of the West result in markedly slower microbiological decomposition, even with equal aeration, than that which occurs in the hotter summers of S.E. England, with its approach to a continental climate.

It is commonly supposed that under conditions favourable for oxidative decomposition, the organic matter status of a soil will fall rapidly in the absence of additions of fresh organic matter. We must distinguish, however, between the organic matter of plant residues and that of humus itself. Plant residues can undergo very rapid decomposition, particularly at high temperatures. Thus, R. K. Tam and O. C. Magistad in Hawaii found that after 35 weeks only 9% of an original incorporation of 18.6 tons of pineapple trash had survived decomposition. Such a rate of decomposition is unlikely in the case of humus which is, for the most part, relatively inert as representing an end-point in microbiological decomposition. F. Scheffer distinguishes between “Nährhumus,” consisting of carbohydrates, organic acids, proteins, and soluble or hydrolysable substances, and “Dauerhumus,” which is resistant to microbial attack. The conditions under which this more resistant humus is built up in the soil are not clearly understood. O. Siegel, for example, found that even long continued manuring with farmyard manure for market garden crops had little effect in increasing this type of humus. On the other hand, the occupation of the soil by gramineous vegetation, or indeed, any
natural vegetative cover, has a favourable effect on organic matter status, presumably owing to the building up of humus reserves. More knowledge is needed concerning the rate at which humus, as distinct from plant residues, undergoes decomposition, under different conditions. Much of the information in the literature relates merely to the decomposition of plant residues.

The effect of changes in land utilization on organic matter status varies considerably; whilst serious depletion in organic matter content through sedative decomposition may occur when virgin soils are brought into arable cultivation, particularly in climates with hot summers, this is not the invariable consequence of agricultural use. Where livestock is kept and considerable proportions of the crops grown are returned in farmyard manure the organic matter status may be stabilized at a level not greatly inferior to that of the virgin soil. L. C. Wheeting¹² found an actual increase in western Washington soils as a result of cropping former forest land. Where forest is replaced by grassland or even by arable culture with grass leys, a rise in organic matter status as compared with forest soil would be expected. The greatest loss of organic matter is likely to occur where steppe or prairie is turned to continuous corn growing as in parts of western America. Here the loss in organic matter may be disastrous in its effect both on plant nutrient status and on soil structure.

NITROGEN CYCLE IN SOILS

The organic matter of soils contains carbon and nitrogen in the approximate ratio of 10:1. Assuming that organic matter contains 55% of carbon and 5.5% of nitrogen, then, if the nitrogenous portion be assumed to be of a protein character, it would amount to $5.5 \times 6.25 = 34.4\%$, or a little over one-third of the total organic matter. There is, however, a distinction between the supposed protein of soil organic matter and fresh proteins, for whilst the latter, on being added to the soil, undergo rapid decomposition, the organic
matter protein appears to be highly resistant to microbiological action, although it may be partially hydrolysed by dilute acids and decomposed by hydrogen peroxide.

There is in all soils a small proportion of soil nitrogen in the form of relatively simple compounds, such as amino-acids, amides, ammonium salts, and nitrates. It is from these compounds, and principally from ammonium salts and nitrates, that plants usually obtain their nitrogen.

It has long been known that nitrates are produced by the decomposition of nitrogenous organic matter. Indeed, until the discovery of the nitre deposits of Chile, the decomposition of organic matter was the main source of the supply of potassium nitrate used for the manufacture of gunpowder. In 1877, T. Schloesing and A. Müntz, under the impetus of the recent discoveries of Pasteur, showed that the production of nitrates from decaying organic matter was due to the action of bacteria. Later work, notably that of R. Warington and S. Winogradsky, showed that the changes involved a successive simplification, according to the following scheme:—Nitrogenous organic matter → amino-acids and amides → ammonium salts → nitrites → nitrates. Whilst the changes from ammonium salts to nitrites, and from nitrates to nitrates were each shown to be the work of specific bacteria, *B. nitrosomonas* and *B. nitrobacter*, respectively, the process of ammonification was found to be the result of the work of many organisms, both fungal and bacterial, and possibly to include purely chemical changes. It should be added that fresh organic matter is much more susceptible to attack than humified organic matter, and is therefore the principal source of ammoniacal and nitrate nitrogen.

Nitrification, the production, via nitrites, of nitrates from ammonium salts, is an aerobic process, the work of autotrophic bacteria whose energy is obtained from the oxidation involved in this change. The minimum, optimum, and maximum temperatures for the process are about 5°C., 25-35°C., and 55°C. respectively. S. P. Tandon and N. R. Dhar conclude that the optimum temperature for nitrification
in tropical soils is about 35°C. against 25°C. in temperate soils. Whilst nitrification proceeds most rapidly in a neutral or slightly acid medium, it is now known that it can also go on, though more slowly, in soils which are definitely on the acid side of neutrality. The presence of moisture is, of course, essential, as for all microbiological actions.

The change, nitrites → nitrates is more rapid than the change, ammonium salts → nitrites; and for this reason nitrites in the free state scarcely occur in soils. The whole change from ammonium salts to nitrates is, under satisfactory conditions for nitrification, more rapid than ammonification. The presence of ammoniacal nitrogen in amounts of more than a few parts per million of soil is evidence that conditions are unfavourable for nitrification. This is commonly the case in acid soils rich in organic matter. Such soils may contain up to 0.05% of ammoniacal nitrogen and be devoid of nitrates. Ordinarily, however, it may be said that the rate of nitrification depends upon the rate of ammonification.

H. L. Richardson has studied the nitrogen cycle in grassland soils at Rothamsted. He found that the content of ammoniacal nitrogen always exceeds that of nitrate nitrogen, but that both are at a low level, which may represent an equilibrium. Injury to herbage by frost or drought causes a rise in the ammoniacal nitrogen level. The effect of added ammonium salts on the ammoniacal nitrogen level quickly disappears.

The production of nitrates in soils is subject to marked seasonal influences. The low temperatures and the deficient soil aeration which obtain in the winter months result in an almost complete cessation of the process. The rate of production increases in the spring, and reaches its maximum in the summer months. The actual proportion of nitrate-nitrogen, even in the same soil, varies enormously, for, not only is it subject to the demands of plant roots, but also, being in the form of soluble salts and not absorbed by the soil, it is rapidly washed out and lost in the drainage, if heavy rainfall occurs. There is also evidence that, apart
from these circumstances, there are rapid fluctuations in nitrate content over short periods.

N. R. Dhar, A. K. Bhattacharya, and N. N. Biswas, and, more recently, G. G. Rao, have suggested that photochemical action may play an important part in nitrification under tropical conditions. Against this, G. S. Fraps and A. J. Sturges found no evidence for photo-nitrification and concluded that the observed effects were due to temperature.

Whilst plants normally absorb their nitrogen in the form of nitrates, certain plants, among them some important crops such as potatoes, appear to prefer ammonium salts. In other cases, the nitrogen nutrition of plants is effected through symbiosis. Certain fungi, notably the Hymenomycetes, are associated with plant roots to form structures known as mycorhiza. In an ectotrophic mycorhiza, the fungal mycelium occurs as a mantle enveloping the root and entering the intercellular spaces, whilst in an endotrophic mycorhiza it is entirely internal. Ectotrophic mycorhizae are commonly found in association with the roots of trees, particularly conifers, for whose satisfactory growth they appear to be necessary, particularly in the seedling stage. M. C. Rayner and W. Neilson-Jones have shown that on poor heath soils in Dorset, the inducement of mycorhizal growth is requisite for the production of healthy seedlings. Mycorhizal combinations are common on acid soils and among plants of semi-xerophytic habit.

The nitrogen absorbed by the plant from the soil, either directly or through the intervention of mycorhiza, is built up into complex plant proteins. These are returned to the soil again as residues, either directly or through the digestive tracts of grazing animals. The extent to which vegetable proteins consumed by human beings are returned as excreta depends on systems of sanitation. Under urban conditions, the nitrogen of human food is almost completely lost.

"Problems in Tree Nutrition" (Faber and Faber).
The cycle, soil nitrogenous matter → ammonium salts → nitrites → nitrates → protein → soil nitrogenous matter is only one of the circulations involved in the complete nitrogen cycle. Apart from the nitrate absorbed by plants, there is a certain loss of nitrate from the soil in drainage. This may be, under British conditions, 10-60 lbs. per acre annually.

There is also the possibility that nitrates may be transformed by bacterial action into elementary nitrogen. Denitrification, as the process is called, represents a definite loss of nitrogen from the soil. It is the work of a bacterium, *B. denitrificans*, the conditions for whose activity appear to be the presence of large quantities of fresh organic matter and the alternation of aerobic and anaerobic conditions. It may be presumed that denitrification is of slight importance except under the highly artificial conditions of market garden soils. P. K. De and S. N. Sarkar\(^{21}\) have shown that added nitrate disappears rapidly under waterlogged conditions and considers this disappearance to be due to assimilation by micro-organisms.

**NITROGEN FIXATION**

We have now to account for the stock of combined nitrogen present in soils. To refer it to the contribution of plant residues is only to change the problem to that of the source of the nitrogen supply of plants. At the outset, we may practically exclude the parent materials of soils as sources of soil nitrogen, since igneous rocks are devoid of this element, whilst sedimentary rocks only contain fractional proportions, which would be quite inadequate to account for the combined nitrogen of soils derived from them.

A possible source is the ammonia and nitric acid brought down by rainfall. This contribution amounts, under British conditions, to about 4 lbs. per acre per annum. Under tropical conditions, however, frequent and intense thunderstorms result in the formation of large quantities of nitric acid. It has been estimated that in Indo-China 30-35
kg. of nitric acid per hectare may fall annually in rainfall. An annual accession of assimilable nitrogen of this magnitude, if shown to be general, might sufficiently account for the nitrogen nutrition of plants and the soil nitrogen of the tropics. Under the conditions obtaining in temperate and cold climates, however, it is necessary to examine other sources of supply.

In the early years of the nineteenth century, the great authority of Liebig was lent to the theory that plants obtained their nitrogen from the traces of ammonia present in the atmosphere, and that the use of nitrogenous fertilizers was as superfluous as the use of fertilizers to supply carbon. The careful experiments of Lawes and Gilbert at Rothamsted, however, clearly demonstrated the dependence of plants on supplies of available nitrogen in the soil. Leguminous plants were an apparent exception, according to field experiments, which actually demonstrated a gain in soil nitrogen consequent on the growth of such crops. In practice, the favourable effect following the growth of leguminous crops had long been known, and is suggested by the familiar lines from Virgil’s First Georgic.

In 1858, Lachmann showed that the nodules which had been observed in the roots of leguminous plants contained bacteria. The problem was cleared up by H. Hellriegel and H. Willfarth in 1888, and the bacteria, finally isolated by M. W. Beijerinck, are now known by the name Bact. radicicola. These bacteria live in symbiosis with their hosts and have the ability to bring atmospheric nitrogen into organic combinations. The nitrogen thus fixed from the air is available for the nutrition of the host plant, from which in turn the bacteria draw their non-nitrogenous pabulum. They are aerobic organisms and appear to exist in forms associated with particular species of leguminous plants.

Certain other plants, notably the alder (Alnus), also exhibit nodule formation and form symbiotic associations with bacteria of the Bact. radicicola group. Symbiotic nitrogen fixation in the genus Casuarina has been studied by H.
Mowry\textsuperscript{25}. V. P. Noglev\textsuperscript{26} reports an organism, \textit{Bacillus alopecuri} (n.sp.) living in symbiosis with the roots of \textit{Alopecurus pratensis} (meadow foxtail) capable of fixing nitrogen. Nitrogen fixation may possibly take place also through mycorhiza. R. Brown\textsuperscript{27} postulates such a fixation by the endophyte of \textit{Lolium}.

In addition to the nitrogen fixation effected by \textit{Bact. radicicola} in symbiosis with leguminous plants, there is another group of bacteria able to build up atmospheric nitrogen into organic combination without the co-operation of a host plant. The free-living nitrogen fixers are heterotrophic organisms depending for their existence on a supply of soluble carbohydrates. The first species, discovered by S. Winogradsky\textsuperscript{28} in 1893, namely, \textit{Clostridium pastorianum}, is an anaerobic organism, originally isolated from pond mud. Under laboratory conditions, it can fix 2-3 mg. of nitrogen per gram of carbohydrate utilized. There is little doubt that bacteria of the \textit{Clostridium} type play an important part in the nitrogen economy of soils with deficient aeration and even of normal soils during periods of deficient aeration.

In 1894, M. W. Beijerinck\textsuperscript{29} discovered an aerobic heterotrophic organism, \textit{Azotobacter chroococcum}, which, like \textit{Clostridium}, utilizes soluble carbohydrates and fixes elementary nitrogen. It appears to be more efficient than \textit{Clostridium} and fixes about 10 mg. of nitrogen per gram of carbohydrate oxidized. \textit{Azotobacter} is more sensitive to acid reaction than \textit{Clostridium} and is generally considered to be inactive in media more acid than pH 6.0. J. K. Wilson and B. D. Wilson\textsuperscript{30}, however, working with New York peat soils, found no relationship between soil reaction and the presence of \textit{Azotobacter}. They consider the nature of the soil complex of more importance than the pH.

Other bacteria are known which can fix atmospheric nitrogen, and symbiotic associations between such organisms and algae have been demonstrated. The direct fixation of nitrogen by higher plants, though occasionally reported, still lacks conclusive proof. F. E. Allison and S. R. Hoover\textsuperscript{31}
report fixation of nitrogen by the blue-green alga, *Nostoc*. P. K. De reports the fixation of nitrogen by algae in Indian rice fields.

During recent years, Indian soil micro-biologists have put forward the possibility that nitrogen fixation may occur by photochemical action. The presence of oxidizable carbohydrate appears to be necessary. Thus N. R. Dhar and E. V. Seshacharyulu found that nitrogen fixation in sugar treated soils was greater in sunlight than in darkness. N. R. Dhar and S. K. Mukherji found that admixture of plant leaves, sawdust, straw, etc., increased nitrogen fixation in sunlight.

**SULPHUR CYCLE IN SOILS**

Sulphur is present in soils mainly in the form of cystine groups associated with the proteins of organic matter. In addition, under anaerobic conditions, sulphides and even free sulphur may be present.

Certain bacteria in the soil have the ability to oxidize organic sulphur compounds, sulphides, and free sulphur to sulphuric acid or sulphates. These are aerobic bacteria whose activity appears to be favoured by the presence of calcium carbonate. Broadly speaking, it may be said that the same conditions which favour nitrification also favour the production of sulphates.

Sulphuric acid from the biological oxidation of organic sulphur compounds may contribute appreciably to the acidity of peat soils. P. Köttgen and Rörig found that from 7.7 to 53% of the total acidity of forest peat soils was due to sulphuric acid. High base-status favours sulphation and may thus accelerate lowering of the pH. Dressings of sulphur are sometimes applied to the soil to increase acidity.

Under anaerobic conditions, sulphates are readily changed into sulphides, including hydrogen sulphide, whose odour is frequently evident in subaqueous horizons.
CARBON-NITROGEN RATIO OF SOIL ORGANIC MATTER

A consideration of the composition of the plant materials added to the soil from the residues of vegetation will show that their ratio of carbon to nitrogen is in the region of 40:1. The organic matter of soils shows a much narrower carbon-nitrogen ratio. For the soils of temperate regions, the ratio approaches 10:1. Only a small proportion of the organic matter in soils consists of unaltered plant residues, and the carbon-nitrogen ratio observed represents the character of the actual soil humus. Sufficient data are not yet available for a satisfactory judgement to be formed as to the effect of climate on the C:N ratio. From a study of soils of the United States, H. Jenny\textsuperscript{36} concludes that the ratio widens with decreasing temperature, but W. McLean\textsuperscript{37} has examined many tropical soils which give C:N ratios considerably greater than 10:1.

M. S. Anderson and H. G. Byers\textsuperscript{38} have examined the application of the C:N ratio to soil classification. Wide variations occur within the groups, the most constant figure, 9.0, being given by the chernozems. The widest ratios, 21.4-22.6, occur in the podsols. It is unlikely that the C:N ratio can give more than a crude indication of the nature of soil organic matter. A wide ratio may be caused by the presence of much undecomposed plant materials or of elementary carbon, the latter forming no inconsiderable contribution in cultivated soils and in soils subject to forest or bush fires.

The addition to soil of organic matter with a wider carbon-nitrogen ratio than that characteristic for the soil is followed by a considerable evolution of carbon dioxide. The organisms concerned in this change require a supply of available nitrogenous compounds and there is thus a temporary diminution in the content of ammoniacal and nitric nitrogen. In the absence of fresh organic matter it may be supposed that losses of carbon and nitrogen from the soil by micro-biological decompositions will be approximately in the proportion in which they occur in the characteristic organic matter of the soil.
THE CONSTITUTION OF SOIL ORGANIC MATTER

The nature and constitution of soil organic matter has long presented a difficult problem for soil chemists. Its complexity was recognised at an early stage by the first workers in the subject. One of the chief difficulties in investigating the problem is the impossibility of isolating the organic matter of ordinary soils from its accompanying mineral matter. It was inevitable, therefore, that much of the earlier work was carried out on peats, consisting wholly or mainly of organic matter. And since the mode of origin of such organic matter is different from that in the majority of ordinary soils, it is probable that the results obtained are only of limited application.

G. W. Robinson and J. O. Jones\(^{39}\) suggested a simple fractionation of soil organic matter into humified and non-humified by the use of boiling 6% hydrogen peroxide as a reagent. The assumption underlying this method of determining the degree of humification was, however, questioned by W. O. Robinson\(^{40}\), who showed that, in the presence of soil, cellulose is appreciably attacked by 6% hydrogen peroxide. The peroxide method for determining the degree of humification of soil organic matter has, however, found a certain application in the study of forest soils. For example, A. Němec\(^{41}\) was able to distinguish different types of forest humus by the use of this reagent.

The fundamental problem of soil organic matter is the nature and mode of origin of the characteristic organic soil constituent, humus. As far back as 1786, F. K. Achard* extracted a black substance from peat by the action of alkali, and this gave the impetus to numerous attempts to isolate complex acids and other related compounds from peats and also from mineral soils. In this way, such compounds as humic acid, ulmic acid, crenic acid, apocrenic acid, humin, and ulmin were distinguished. But, inasmuch as they are non-crystalline substances of exceedingly complex composi-

tion, their identity as definite chemical compounds has always been open to doubt.

A possible fractionation of soil organic matter is shown in the following scheme:

```
Organic Matter

Sol. in cold alkali  Insol. in cold alkali (humin)

Sol. in water or dil. acids (apocrenic acid)

Precip. by dil. acids

Sol. in water (Fulvic acid)  Sol. in alcohol (hymatomelanic acid)  Insol. in alcohol (humic acid)
```

The very existence of humic and similar acids was called into question by certain workers, notably E. Gully, who considered that the phenomena of extraction by alkalies and precipitation by acids were explainable as colloidal absorption effects.

The existence of acids in peat was placed beyond doubt by the work of S. Odén, who showed that the conductometric curves obtained by adding increasing amounts of alkali to peat or prepared humic acids were such as could only be given by weak insoluble acids.

Odén concluded that humic acid has the formula $C_{40}H_{52}O_{24}(COOH)_4$. Its aqueous suspension has a pH of 4.5. It should be remarked, however, that pH values of less than 4 are frequently encountered in podsol profiles under heath peat.

O. Schreiner and E. C. Shorey succeeded in isolating a number of definite compounds from soil organic matter. Included in the humic acid fraction, they found paraffinic acid, agroceric acid, $C_{26}H_{44}O_3$, agrosterol $C_{26}H_{44}O$, and phytosterol, $C_{26}H_{44}O$. In the fraction not precipitated by acids (crenic and apocrenic group) they found dihydroxystearic acid, $\alpha$-picoline-$\gamma$-carboxylic acid, xanthine,
hypoxanthine, cytosine, histidine, and arginine. No complete fractionation was effected and it is possible that some of the compounds isolated were not originally present as such in the soil, but were formed by the processes of extraction used by these workers.

It has long been known that, by the action of strong hydrochloric acid on sugar, various substances similar in appearance to humus can be obtained, and it was supposed that they might be similar in origin and constitution. Artificially prepared humic acid is, of course, free from nitrogen, whereas natural humic acid always contains nitrogen; but it was thought possible that such nitrogen might be present in adventitious compounds which were not essential constituents of humus. It was, however, found possible to prepare artificial humic acids containing nitrogen by condensation of sugars with amino-acids. L. C. Maillard actually suggested this mode of origin in the soil.

V. H. Beckley demonstrated the formation of humus from carbohydrates by way of furfuraldehyde or methylhydroxyfurfuraldehyde, which undergo condensation to form humus. W. Eller suggested an entirely different mode of origin by way of benzene derivatives. Humic acids prepared artificially by the oxidation of phenol derivatives were shown to resemble closely the humus of soil.

The fact that humus has a dark colour has tended to obscure the possibility that a considerable proportion of it may consist of materials which have survived from plant residues. R. A. Gortner found that alkaline extracts similar to those obtained from soils could also be obtained from unchanged vegetable materials. He suggested that the only characteristic soil constituent is a black pigment, probably of bacterial origin, which rarely exceeds 40% of the humus obtained by alkaline extraction. The pigment, which is poor in nitrogen, is soluble in aqueous ammonia.

During recent years interest has centred round the origin of humus from the lignin and protein of plant residues. Starting from the suggestion of F. Fischer and H. Schrader...
that lignin was the mother-substance of humus, S. A. Waksman and his co-workers have conclusively demonstrated the important rôle of this substance in the transformations of soil organic matter. Thus, in 1926, Waksman suggested that the alkali-soluble humus of soils consists of two main categories, namely, (a) lignin, which being unattacked by micro-organisms, persists from added plant materials; and (b) microbial protoplasm and its decomposition products. The substances in the second group contain nitrogen. M. S. Du Toit and H. J. Page found a correlation between loss of lignin and gain in humus.

S. A. Waksman and K. R. Stevens proposed the application to soil organic matter of a system of proximate analysis used for plant materials. The results for the organic matter of some typical soils are shown in Table XIII. The ether-soluble fraction includes essential and fatty oils, wax- and resin-like substances, nitrogenous fats, etc. The alcohol-soluble fraction includes waxes, resins, alkaloids, tannins, and chlorophyll. Hemicellulose includes pentosans and hexosans. Cellulose includes celluloses and chitin-like substances. Lignin-humus is given by difference.

**Table XIII.—Composition of organic matter for some typical soils. Calculated on basis of soil organic matter = C × 1.72 (Waksman and Stevens)**

<table>
<thead>
<tr>
<th>SOIL</th>
<th>Ether Soluble</th>
<th>Alcohol Soluble</th>
<th>Hemi-Celulose</th>
<th>Cellulose</th>
<th>Lignin Humus Complex</th>
<th>Protein</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summit soil, 4.49% org. matter...</td>
<td>3.56</td>
<td>0.58</td>
<td>5.44</td>
<td>3.55</td>
<td>43.37</td>
<td>33.78</td>
</tr>
<tr>
<td>Kansas chernozem, 5.98% org. matter</td>
<td>4.71</td>
<td>1.53</td>
<td>8.60</td>
<td>5.22</td>
<td>40.81</td>
<td>34.74</td>
</tr>
<tr>
<td>Alberta chernozem, 11.20% org. matter</td>
<td>0.80</td>
<td>0.82</td>
<td>5.53</td>
<td>4.12</td>
<td>41.87</td>
<td>37.35</td>
</tr>
<tr>
<td>Carrington loam, 5.48% org. matter</td>
<td>0.62</td>
<td>0.62</td>
<td>8.21</td>
<td>3.64</td>
<td>42.29</td>
<td>30.38</td>
</tr>
</tbody>
</table>
Compared with plant residues such as straw and leaves, the soil organic matter contains much less of the hemicellulose and cellulose fractions. On the other hand, there is a considerable increase in the lignin fractions and the protein fraction. It appears unlikely that the lignin shown in the above system of fractionation is identical with plant lignin, whilst the identity of organic matter protein with plant protein is, as will be seen later, also improbable. Whatever be the nature of these fractions, they appear to account for 75% of the total organic matter of the soil.

The peats present a special type of distribution among the different fractions. Whilst in fen peats, cellulose is almost absent, in high moor (sphagnum) peats, considerable proportions of cellulose and hemicelluloses appear to survive decomposition. Protein is considerably higher in low moor than in high moor peat.

N. P. Remesnov has examined different kinds of soil by Waksman's methods and finds important differences. Thus, celluloses and hemicelluloses are absent from grey soils and protein predominates over lignin humus. In chestnut earths hemicelluloses appear, whilst lignin humus is greater than protein. In chernozems the lignin humus is 2-5-3 times the protein.

It soon became evident that both the lignin and the protein of humus differ from the original lignin and protein of plant residues. For example, W. T. McGeorge showed that humus lignin possesses a higher base exchange capacity than plant lignin. According to H. Pallmann, humic acid shows a higher exchange capacity, probably due to increase in the proportion of carboxyl groups, which are absent from lignin, and a higher degree of oxidation as shown by the O:C ratio. It may be intermediate in properties between lignin and graphitic acid. Further, the protein of humus differs markedly from the protein of plant materials in being resistant to nitrification. R. P. Hobson considered the "humic acid complexes" to be adsorption compounds of lignins and proteins.
Evidence for regarding humus as a single complex was obtained by W. McLean from a study of the attack on soil organic matter by hydrogen peroxide. Whilst it is undoubtedly true that 6% hydrogen peroxide, originally proposed for differentiating humified from non-humified or structural matter, has an appreciable attack on cellulose in the presence of soil, a weaker peroxide (1-2%) is without action on this substance. By studying the effect of hydrogen peroxide of varying strength on the carbon and nitrogen content of soils, it was found that two stages of oxidation could be distinguished. Starting with the most dilute solutions, a complex of constant carbon-nitrogen ratio is progressively oxidized. Above a certain strength of peroxide, varying somewhat for different soils, there is no further oxidation of nitrogenous matter, but a progressive decomposition of non-nitrogenous compounds or complexes. The course of the oxidation of a Welsh brown earth is shown graphically in Fig. 10. It will be seen that, even with very dilute peroxide, there is a considerable oxidation of carbon and nitrogen; but whilst the oxidation of carbon slowly increases after about 86% has been attacked, there is no further attack of nitrogen beyond the first stage of attack.

The material oxidized in the first stage, termed by McLean the oxidisable complex, accounts for 70 to 80% of the total carbon and nitrogen present, and is more readily oxidized in base-unsaturated than in base-saturated soils. It would appear to be identifiable with the lignin-humus complex and the protein of Waksman (loc. cit.). The non-nitrogenous organic matter oxidized in the second stage may represent cellulosic material which has not as yet undergone microbiological decomposition.

Whilst, according to Waksman's earlier ideas, the lignin-humus and the protein were regarded as distinct complexes, the one representing resistant material surviving microbiological decomposition and the other consisting mainly of altered microbial protoplasm, the constancy of the carbon-nitrogen ratio of the organic matter oxidized by weak
peroxide does imply some close association between these groups. The high resistance of humic nitrogenous compounds to attack by ammonifying organisms compared with that of known plant and animal proteins suggests that the nitrogen of the oxidizable complex may be built up in com-

![Graph](image-url)

**Fig. 10.** Oxidation of soil organic matter by hydrogen peroxide.

plexes such as the artificial "humus" prepared by L. C. Maillard (*loc. cit.*) from the condensation of carbohydrates and amino-acids.

More recently, S. A. Waksman and K. R. N. Iyer have actually demonstrated the formation of a complex, from the combination of lignin and protein, which appears to be similar to soil humus in its properties. To this they apply the term *humus-nucleus*. There is good evidence for believing the humus of soils to be mainly of this character and to originate from the combination of plant lignins with
proteins. Evidence for its existence as a definite though complex compound was obtained by I. Siedlecky and B. Brunowski\textsuperscript{69}, who found all humus preparations to be crystalline and to give the same Debye-Scherrer diagram when examined by X-ray analysis. W. S. Gillam\textsuperscript{69} found humic acid from three soil types to be chemically and physically similar.

It is necessary to emphasize the difference which appears to exist between plant lignin and the combined lignin of humus. Reference has already been made to McGeorge's evidence that soil lignin has a greater exchange capacity than plant lignin. It was considered originally by Waksman that lignin was resistant to biological decomposition. The rapid disappearance of plant residues under certain conditions, however, suggests its susceptibility to change. Further, M. Phillips, H. D. Weihe, and N. R. Smith\textsuperscript{69} have shown that under certain conditions lignin may be as rapidly decomposed by soil organisms as celluloses and proteins. A. G. Norman\textsuperscript{61} found 40-50\% of the lignin of oat straw to be removed in one year by aerobic decomposition at 28\textdegree C.

According to the foregoing ideas, the characteristic constituent of soil organic matter, humus, is a compound of altered lignin with proteins originating from plant proteins and the protoplasm of soil organisms, occurring as a single homogeneous ingredient, which, however, may be expected to have differing composition in different types of soils. The small amounts of celluloses and hemi-celluloses present must be regarded as fugitive in character.

As against the view that humus is a residual product of the decomposition or alteration of lignin, F. Y. Geltser\textsuperscript{62} considers that humic acid is formed from the products of the decomposition of plant residues by a distinct group of syntheses and that it is not necessarily related in constitution to the constituents of these residues. He postulates two types of humus, one characteristic of chernozems and cultivated soils, formed by the action of bacteria, and the other formed by fungal action occurring in peats and podzols. P. H.
Gallagher distinguishes the “mobile” humus in podsols as a separate type, differing from other types of humus by its solubility in oxalic acid and by its low nitrogen content.

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humus" compounds and of natural humus upon soil microbiological processes. III. The base exchange capacity of synthesized humus and of natural humus complexes. IV. Fixation of lignin by proteins and formation of complexes resistant to microbial decomposition. *Soil Sci.*, 1932, 34, pp. 43-70; 1933, 36, pp. 57-67; 69-82


CHAPTER VIII

GENERAL PHYSICAL PROPERTIES OF SOILS

Soils exhibit certain peculiarities in their physical properties by reason of the fact that they are essentially mixtures of mineral particles of varying size, together with closely associated colloidal clay and colloidal organic matter. The physical character and behaviour of a soil are governed by a number of circumstances, namely:

(1) The nature and the size-distribution of the particles composing the non-colloidal skeleton or framework.

(2) The content of colloidal material.

(3) The character of the colloidal material, depending on (a) the relative proportions of organic and inorganic colloidal material; (b) the constitution of the inorganic colloidal material; and (c) the nature and content of the exchangeable bases present.

(4) The structure of the soil, i.e. the extent to which the primary particles of the soil are built up into aggregates, and the character of these aggregates or structural elements.

(5) The moisture content.

TRUE AND APPARENT DENSITY

It is necessary to distinguish, in the first place, between the true and the apparent density of a soil. Whilst the true density is an additive function of the densities of the soil constituents, the apparent density will vary according to the state of aggregation of the soil and the proportion of the apparent volume occupied by the interstitial spaces which are present even in the most compact soil.

The true density of a soil depends principally on the relative proportions of inorganic and organic matter present;
for, apart from certain heavy minerals, such as magnetite, which rarely occur in appreciable quantities, the densities of the inorganic constituents of the soil fall within a fairly narrow range. The following are the densities of the principal mineral ingredients of soils:

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Density</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>2.6-2.7</td>
</tr>
<tr>
<td>Orthoclase</td>
<td>2.54-2.57</td>
</tr>
<tr>
<td>Albite</td>
<td>2.605</td>
</tr>
<tr>
<td>Oligoclase</td>
<td>2.65</td>
</tr>
<tr>
<td>Labradorite</td>
<td>2.68-2.71</td>
</tr>
<tr>
<td>Anorthite</td>
<td>2.765</td>
</tr>
<tr>
<td>Apatite</td>
<td>3.16-3.22</td>
</tr>
<tr>
<td>Magnetite</td>
<td>4.9-5.2</td>
</tr>
<tr>
<td>Amphiboles &amp; Pyroxenes</td>
<td>2.9-3.6</td>
</tr>
<tr>
<td>Olivine</td>
<td>3.2-3.5</td>
</tr>
<tr>
<td>Muscovite</td>
<td>2.76-3.1</td>
</tr>
<tr>
<td>Biotite</td>
<td>2.8-3.2</td>
</tr>
<tr>
<td>Chlorite</td>
<td>2.6-2.96</td>
</tr>
<tr>
<td>Haematite</td>
<td>4.9-5.3</td>
</tr>
<tr>
<td>Limonite</td>
<td>3.6-4.0</td>
</tr>
<tr>
<td>Clay</td>
<td>2.6-2.7</td>
</tr>
</tbody>
</table>

The density of soil organic matter varies from 1.2 to 1.7, but the figure is open to some doubt since the organic matter of ordinary soils cannot be isolated unchanged.

Since by far the greater part of the mineral matter of ordinary soils consists of quartz, felspars, and clay, the density of the mineral soil free of organic matter is generally about 2.65. It is lowered by the presence of organic matter, and raised by the presence of oxides of iron such as magnetite, haematite, and limonite, and, to a smaller degree, by the presence of ferromagnesian minerals and micas. It is generally assumed that the density of a soil is an additive function of the densities of its constituents.

The true density of a soil has proved of comparatively little value for diagnostic purposes, since the presence of notable proportions of organic matter or hydrated ferric oxide is readily detected by other means. It must be admitted, however, that no attempts have been made to correlate density with soil constitution over the very wide range of soil which it is necessary to consider in modern investigations. If such correlations were worked out, the density might prove to have a certain significance and serve as an aid to diagnosis where other methods of investigation are impracticable.

Since the density of solid materials is generally obtained by methods depending on the use of liquids of known den-
sity, it follows that the density, and hence the specific volume, actually found will be affected by any surface reactions between the solid material and the liquid used. M. S. Anderson and S. Mattson found that the specific volume of clays was highest in non-polar* liquids such as carbon-tetrachloride and lowest in polar liquids such as water. E. W. Russell has shown that this may be attributed to the drawing together of the molecules of polar liquids in the electrostatic fields of force associated with exchangeable ions, and has deduced formulæ to express this effect quantitatively. According to these ideas, the true specific volume of a clay should be regarded as the value obtained in a non-polar liquid.

PORE-SPACE +

If a certain volume of a soil in its natural conditions be considered, it is evident that only a proportion of that volume is occupied by the soil material. The remainder consists of interstitial spaces which, under ordinary field conditions, are occupied partly by water and partly by air. The weight of unit volume of dry soil, including the interstitial spaces, gives the apparent density.

The relationship of the true density, \( d \), and the apparent density, \( d_1 \), to the pore-space will be apparent from the following considerations. If the percentage of pore or interstitial space be \( P \), then the true volume of 100 c.c. of soil will be 100-\( P \), and the weight \((100-P)d\). The apparent density will therefore be \((100-P)d/100\), or \(d-Pd/100\). The pore-space expressed in terms of true and apparent density is given by \(P = 100(d-d_1)/d\).

*By a polar liquid is understood one in whose molecules the electrons are assymetrically arranged round the nuclei, i.e., in which the electrical centre of gravity of the electrons is not coincident with the electrical centroid of the nuclei. In a non-polar liquid the distribution of the electrons round the nuclei of the molecules is symmetrical. In the former case the molecules will tend to orient themselves in an electrostatic field.

+The subject matter of this section has already appeared in Chapter II, but is given again here for the sake of completeness.
F. H. King\textsuperscript{3} gives the following figures for the percentage pore-space of typical soils:

<p>| | | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Sandy soil</td>
<td>32.5</td>
<td>Loamy clay</td>
</tr>
<tr>
<td>Loam</td>
<td>34.5</td>
<td>Clay loam</td>
</tr>
<tr>
<td>Heavy loam</td>
<td>44.1</td>
<td>Heavy clay</td>
</tr>
</tbody>
</table>

Such figures should be regarded only as giving orders of magnitude, for wide variations may occur in the same soil with changes in aggregation. Pore-space is increased by addition of organic matter, and may rise in grass soils to 60\%, or more. Clay soils show wider variation in pore-space than sands.

Taking 50\% as the pore-space under natural conditions, it is apparent that this does not represent the closest possible packing of the particles of which the soil is composed. With uniform spheres, the tightest packing gives a pore-space of about 26\%. Where particles of different size are present and can pack into the interstices between the larger particles, the pore-space, theoretically, can be considerably reduced. It is evident that, with the particle-range of ordinary soils, the tightest possible packing would result in a pore-space very much less than 50\%. The high figure actually obtained is due to the aggregation of particles into compound particles or crumbs. Any treatment which destroys the crumb structure and results in the soil particles behaving singly leads to a decrease in pore-space. This change may be brought about mechanically by trampling, or by the beating of heavy rain, and chemically by the action of certain substances known as deflocculants, of which sodium carbonate is the chief example. The change of a saline soil into an alkaline soil consequent on the removal of soluble sodium salts is accompanied by deflocculation, involving decrease in pore-space.

Pore-space generally decreases from the surface downwards, but may pass through a minimum in an illuvial horizon, particularly if there is compaction. Unfortunately, its determination under field conditions is beset with many difficulties.
It is clear that the pore-space in a soil varies both in dimensions and form. It may be regarded as forming a continuous three-dimensional network of channels and spaces of varying size. Just as we may speak of a particle-size distribution we may also speak of a pore-size distribution. Some writers have attempted to distinguish between capillary and non-capillary pore-space, the latter being considered as consisting of those voids that are occupied with air when the soil is moistened up to the capacity implied by the moisture equivalent. Such a distinction, although useful, is arbitrary and it is more correct to think of pore-space distribution as a continuous function. The form of this distribution has been investigated by determining the amount of moisture retained by the soil at varying suction pressures. F. Sekera⁴ has calculated the equivalent pore diameter for different suction forces as follows:

<table>
<thead>
<tr>
<th>Suction force (atm.)</th>
<th>0.03</th>
<th>0.1</th>
<th>0.3</th>
<th>1.0</th>
</tr>
</thead>
<tbody>
<tr>
<td>Equivalent diameter</td>
<td>100μ</td>
<td>30μ</td>
<td>3μ</td>
<td>10μ</td>
</tr>
</tbody>
</table>

The equivalent diameter is thus inversely proportional to the suction pressure. The free capillaries in which water can circulate depend on the immobile film water. Coarse capillaries are defined as those greater than 30μ, fine capillaries as those less than 3μ.

**TOTAL SURFACE OF SOILS**

The subdivision of the soil material implied by its granular character involves a great development of surface, which is the more marked the finer is the texture of the soil. This is clear from the following considerations.

A cube having sides 1 cm. in length has a total surface of 6 cm². If, now, the cube be subdivided into millimetre cubes, the total surface will become $1000 \times 6(0.1^2)$ cm², or 60 cm². The general relationship among particles of similar shape is that the specific surface, i.e., the surface per unit weight, is inversely proportional to the linear dimensions, i.e., in the case of spheres, inversely proportional to the radius.
GENERAL PHYSICAL PROPERTIES OF SOILS

It is possible, from a consideration of the mechanical composition curves of soils, to obtain an estimate of their specific surface.

The total surface of unit weight of a uniform soil fraction of diameter \( D \), assuming the particles to be spherical, is equal to:

\[
\text{Number of particles} \times \text{surface of each particle} = \frac{1}{v d} \times \pi D^2
\]

[Where \( v \) = volume of each particle, \( d \) = density, for which the value 2.65 may be assumed, \( D \) = diameter of particles.]

\[
= \frac{1}{\pi D^3} \times \pi D^2 = \frac{2.264}{D}
\]

If 100 g. of soil consist of \( p_1, p_2, p_3 \ldots p_n \) per cent. of particles having diameters \( D_1, D_2, D_3 \ldots D_n \), respectively, the total surface is equal to

\[
2.264 \left( \frac{p_1}{D_1} + \frac{p_2}{D_2} + \frac{p_3}{D_3} + \ldots \frac{p_n}{D_n} \right)
\]

A measure of the total surface can be obtained graphically by the method of G. Krauss\(^a\) depending on the following argument.

Let the composition of the soil be set out in the form of a number of sectors, in which the angle of each sector is proportional to the percentage, \( p \), of each fraction, and the radius proportional to \( 1/\sqrt{v} \), the reciprocal of the fourth root of the settling-velocity.

Since the area of a sector, \( S \), is proportional to the angle, and to the square of the radius,

\[
S = k_1 p / \sqrt{v} \quad \text{(} k_1 \text{ = a constant)}
\]

But, from Stokes' Law, \( D \propto \sqrt{v} \),

\[
\therefore \quad S = k_2 p / D \quad \text{(} k_2 \text{ = a constant)}
\]
The area of unit angle of each sector is thus inversely proportional to the diameter and, consequently, proportional to the total surface of the fraction represented by it; and the sum of the areas of the sectors is proportional to the total surface of the soil.

In view of the smooth character of the mechanical composition curves of soils, it is sufficient to mark off at angular intervals, equivalent to 10%, radii proportional to the reciprocal of the fourth root of the corresponding settling
velocity, beginning with the lowest settling velocity. The area obtained by joining up the extremities of the radii is evaluated by means of a planimeter.

Curves drawn in this manner are shown in Figs. 11 and 12. The data are taken from the mechanical composition curves shown in Figs. 1 and 2.

![Diagram of soil composition](image)

**Fig. 12.**—Total surface of certain soils.

The areas in square centimetres on the scale used and the estimated surfaces are shown in Table XIV. The latter figures are calculated on the assumption that the total surface per gram of uniform spherical particles, 0.002 mm. in diameter, is $2.264/0.0002 = 11,320$ square centimetres, and
that such particles have a settling-velocity equivalent to 10 cm. in 8 hours.

Inspection of the curves will show at once the small total surface possessed by mechanically comminuted materials such as slate powder and quartz powder, even although they contain very little coarse material. The small contribution of the coarser fractions to the total surface is also evident.

**Table XIV.—Total Surface of Typical Materials**

<table>
<thead>
<tr>
<th>Material</th>
<th>Area included by curve sq cm.</th>
<th>Estimated surface, square cm. per gram.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz powder</td>
<td>...</td>
<td>1.16</td>
</tr>
<tr>
<td>Kaolin</td>
<td>...</td>
<td>7.80</td>
</tr>
<tr>
<td>Slate powder</td>
<td>...</td>
<td>15.87</td>
</tr>
<tr>
<td>Shaly light loam, Caernarvon...</td>
<td>...</td>
<td>18.45</td>
</tr>
<tr>
<td>Sandy red loam, East Africa...</td>
<td>...</td>
<td>36.97</td>
</tr>
<tr>
<td>Clay subsoil, Rothamsted...</td>
<td>...</td>
<td>43.48</td>
</tr>
<tr>
<td>Raw Carboniferous clay, Flint</td>
<td>...</td>
<td>60.31</td>
</tr>
<tr>
<td>Clay separate, Sudan</td>
<td>...</td>
<td>118.5</td>
</tr>
<tr>
<td>Heavy red loam, East Africa...</td>
<td>...</td>
<td>164.4</td>
</tr>
<tr>
<td>Uniform fraction, 0.002 mm. diameter (calculated)</td>
<td>...</td>
<td>6.86</td>
</tr>
</tbody>
</table>

The estimates thus obtained are probably no measure of the actual specific surface of the soils to which they refer, since under natural conditions a large proportion of the colloidal material, mainly responsible for these large figures, is probably built up into loose structures that become broken down in the operations of analysis. The values obtained may, however, serve a useful purpose in giving a comparative view of the degree of dispersion of which the specific surface is a measure.

From measurements of hygroscopicity, E. A. Mitscherlich obtained values for the specific surface, varying from 138 cm.² for fine sand to 96,670 cm.² for a clay from Java. These are now generally agreed to be over-estimates.
HEAT RELATIONSHIPS OF SOILS

The temperature of the soil is closely related to the temperature of the air. But, whilst the mean temperature of the soil throughout the year approximates very closely to the mean temperature of the air, the seasonal and diurnal variations show different courses according to the depth below the surface. Below a certain depth, the temperature of the subsoil is no longer affected by seasonal and diurnal changes and shows the ordinary increase with depth due to the influence of the earth's internal heat.

The response of the soil to additions or losses of heat in the form of radiation depends mainly on its specific heat, but also to some extent on losses of heat through vaporization from wet soil. Whilst the specific heat of dry soil shows a certain amount of variation, far greater variations are due to changes in moisture content, owing to the high specific heat of water. The effect of varying moisture contents on the specific heat of soils is illustrated by the data from E. A. Mitscherlich in Table XV.

**Table XV. — Volume specific heat (calories per c.c.) of soils with varying moisture contents (Mitscherlich)**

<table>
<thead>
<tr>
<th>Soil</th>
<th>Volume specific heat in calories per c.c. at varying degrees of saturation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Dry</td>
</tr>
<tr>
<td>Sand</td>
<td>0.302</td>
</tr>
<tr>
<td>Humus</td>
<td>0.148</td>
</tr>
<tr>
<td>Clay</td>
<td>0.240</td>
</tr>
</tbody>
</table>

In considering the specific heat of soils, it is generally desirable to express it on a volume rather than on a weight basis. From the point of view of field conditions, this leads to a truer estimate of the response of the soil to gains or losses by radiation than that obtained by the ordinary method of expression.
Soils that are habitually moist respond more slowly to the heating effect of solar radiation than dry well-drained soils. And thus, in the spring, the beginning of active plant growth takes place earliest in the driest soils.

The character of the diurnal variations in soil temperature is seen from the following data obtained by Mutterich in a sandy soil in Eberswald, Germany, in June, 1889. (Table XVI.)

**Table XVI. — Air temperature, and soil temperature at varying depths (Mutterich)**

<table>
<thead>
<tr>
<th>Time</th>
<th>Temperature °C.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Air</td>
</tr>
<tr>
<td>12 midnight</td>
<td>13.80</td>
</tr>
<tr>
<td>2 a.m.</td>
<td>12.90</td>
</tr>
<tr>
<td>4 a.m.</td>
<td>12.53</td>
</tr>
<tr>
<td>6 a.m.</td>
<td>14.68</td>
</tr>
<tr>
<td>8 a.m.</td>
<td>17.99</td>
</tr>
<tr>
<td>10 a.m.</td>
<td>21.05</td>
</tr>
<tr>
<td>12 noon</td>
<td>21.97</td>
</tr>
<tr>
<td>2 p.m.</td>
<td>22.61</td>
</tr>
<tr>
<td>4 p.m.</td>
<td>22.38</td>
</tr>
<tr>
<td>6 p.m.</td>
<td>21.24</td>
</tr>
<tr>
<td>8 p.m.</td>
<td>17.55</td>
</tr>
<tr>
<td>10 p.m.</td>
<td>14.80</td>
</tr>
<tr>
<td>Mean</td>
<td>17.79</td>
</tr>
</tbody>
</table>

The actual temperatures observed under other conditions will vary according to the course of the air temperature, the intensity of the solar radiation during the day, and the rate of loss by radiation during the night; yet the data in the above table illustrate the main features of the temperature changes in the soil. In sunny weather, the temperature of the actual surface of bare soil reaches a higher maximum than the air. Indeed, surface temperatures as high as 85°C have been reported from the tropics. Below the surface, the
Eberswald data show that, with increase of depth, the amplitude of the diurnal temperature changes decreases, whilst the maxima and minima of the curve show an increasing lag behind the maxima and minima on the air temperature curve. Thus, whilst in the air, the maximum is at 2 p.m., and the minimum at 4 a.m., at 15 cm., the maximum is at 4 p.m. and the minimum at 6 a.m., and at 60 cm., the maximum is at 6 a.m. and the minimum at 6 p.m. The corresponding amplitudes are 7.81°, 5.56°, and 0.17°, respectively.

Under a cover of grass vegetation, the amplitude of the temperature changes at the surface is less than in bare soil, but there is a tendency for the minimum to be below that in the air. Still smaller variations occur under forest. But in every case there is the same tendency for amplitude to decrease with depth and for the changes to lag behind those at the surface.

I. de V. Malherbe⁹ found in S. Africa that where the annual air temperature amplitude was 66-70°F., the corresponding figures in the soil were at 3", 58.6°, at 6", 39.2°, at 12", 28.8°, and 24", 24.6°. The mean daily range at 3" decreased from 23.4° in January to 3° in July. The mean range at 24" was 0.7-2°. Summer rain lowers temperature at 3" by about 15°.

Situation and aspect affect the temperature of the soil. For each 300 feet increase in altitude there is a fall of approximately 1°F in air temperature, which is reflected also in the soil temperature. South slopes, in the Northern Hemisphere, are warmer than north slopes since they receive more intense isolation. (See also p. 109.)

Colour affects soil temperature. Dark-coloured and red soils absorb solar radiation better than light-coloured soils. Wollny, comparing the temperature, at 4 inch depths, of soils covered with black and white material, respectively, found differences of nearly 6°C in favour of the former.

The seasonal changes in temperature show a similar tendency. With increase in depth, the difference in mean
temperature between summer and winter becomes smaller and finally disappears. From A. A. Rambaut's\textsuperscript{19} observations at Oxford, it appears that, under a grass vegetation, the temperature at 10 feet varies from a minimum of about 45.5°F in March to a maximum of 56°F in September; at 6 feet, the range is from rather less than 44°F in February to 59°F in August; whilst at 6 inches, the range is from 38.5°F in February to 66.5°F in July. It is thus evident that the temperature of the soil lags markedly behind that of the air, and that lag increases and amplitude decreases with the depth.

Whilst variations in soil temperature are due mainly to the seasonal and diurnal changes in solar radiation, minor fluctuations are caused by other circumstances. Clouds check the loss of heat from the soil by radiation during night and hinder access of heat by radiation during day. Drying winds blowing over moist soils cause a lowering of temperature corresponding with the latent heat required for the evaporation of moisture. Precipitation on cold soil from moist warm air results in gains of heat.

The freezing of soils takes place normally from the surface downwards. In regions with perpetually frozen subsoil there is thus at the onset of winter a zone generally saturated with liquid moisture lying between the frozen surface and the frozen subsoil. In Britain where the temperature at depth tends to approximate to the mean animal temperature, about 50°F., frost rarely penetrates more than a foot or so. Even in the more continental climate of France according to H. Geslin\textsuperscript{12}, the depth of freezing rarely reaches one metre. V. Novak\textsuperscript{13} in Czechoslovakia found that penetration was less under grass. Compaction increases penetration. Whilst in arable, penetration was rarely more than 80 cm., under field roads it reached 100 cm. and under town roads, 160 cm. Geslin (loc. cit.) also states that compaction increases thermal conductivity, and presumably the rate of penetration of frost. Free water, i.e., the water in the larger capillaries, freezes first and the ice crystals thus formed grow at the expense of the moisture in the finer capillaries.
The "heaving" effect of frost depends on the state of compaction and on the moisture content. A snow cover generally stabilizes temperature and hinders penetration of frost. Whilst thawing generally takes place from the surface, according to R. K. Schofield it may take place from below when snow falls or where an initially severe frost moderates.

SHRINKAGE AND EXPANSION

Anyone familiar with a clay country will have noticed the cracking of the surface of the soil during drought. The shrinkage on drying which leads to this cracking is most marked in the heaviest clays. The fissures thus opened often extend several feet into the soil and result in a further intensification of the drying-out process. Material from the surface sometimes falls into the fissures and it has been shown by H. Greene\textsuperscript{11} that, in the Sudan Gezira, this leads to a considerable circulation of soil material.

Shrinkage on drying is also observed in highly organic soils. The drying-out consequent on the drainage of peat soils is accompanied by marked contraction, so that the surface may be appreciably lowered.

Under laboratory conditions shrinkage is studied by moulding soil with water into a plastic paste and measuring the volume occupied at varying moisture contents during drying. In such a paste, the volume is represented by the soil and water only, for, by the method of preparation, interstitial air spaces, such as occur under natural conditions, are excluded. The curve connecting volume with moisture content is at first linear, the decrease in volume corresponding exactly with the loss in moisture. As the paste becomes drier, however, the curve begins to depart from the initial straight line and the diminution in volume lags behind the loss in moisture. At this point interstitial air spaces make their appearance and there is a change in colour towards that of the dry soil. The shrinkage that occurs after the linear relationship has been departed from is termed residual shrinkage.
W. B. Haines has determined the residual shrinkage of materials formed into plastic blocks. His results are shown in Table XVII.

**Table XVII. — Shrinkage of Plastic Blocks and Clays (Haines)**

<table>
<thead>
<tr>
<th>Soil</th>
<th>Total Shrinkage</th>
<th>Residual Shrinkage</th>
<th>Clay % (log v = 4)</th>
<th>Ignition Loss %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay separate</td>
<td>130.0</td>
<td>8.6</td>
<td>90.5</td>
<td>9.5</td>
</tr>
<tr>
<td>Kaolin</td>
<td>30.0</td>
<td>nil</td>
<td>52.8</td>
<td>12.4</td>
</tr>
<tr>
<td>Sudan clay</td>
<td>67.5</td>
<td>7.6</td>
<td>42.3</td>
<td>11.4</td>
</tr>
<tr>
<td>Clay subsoil</td>
<td>34.0</td>
<td>2.0</td>
<td>33.8</td>
<td>9.6</td>
</tr>
<tr>
<td>Harpenden common</td>
<td>37.5</td>
<td>10.3</td>
<td>25.5</td>
<td>10.5</td>
</tr>
<tr>
<td>Rothamsted loam</td>
<td>21.0</td>
<td>2.2</td>
<td>14.8</td>
<td>8.3</td>
</tr>
<tr>
<td>Sandy soil</td>
<td>9.5</td>
<td>2.6</td>
<td>5.3</td>
<td>8.3</td>
</tr>
<tr>
<td>Peat soil</td>
<td>85.5</td>
<td>18.2</td>
<td></td>
<td>27.0</td>
</tr>
</tbody>
</table>

Whilst the total shrinkage represents roughly the water holding capacity of each material, the residual shrinkage appears to be connected with the presence of colloidal gel material present. It varies from nil in the case of kaolin to 18.2% in the case of a peat soil. In the case of kaolin or any other gel-free powder, shrinkage corresponds with water loss until the grains come into contact, after which no further shrinkage can occur. But where gel material is present, it is interposed between the mineral grains and opposes a resistance to shrinkage in the later stages of drying. The property of residual shrinkage can be conferred on non-colloidal powders such as kaolin by admixture of colloidal material such as gelatin.

The results obtained in the laboratory are only partially representative of field conditions, where, even in wet soils, a certain proportion of air spaces is always present above ground-water level. The laboratory conditions are most closely simulated by the individual "clods," which, in varying proportions, are always present in clay soils.
GENERAL PHYSICAL PROPERTIES OF SOILS

The shrinkage of soils on drying produces certain characteristic structures such as prismatic, columnar, and nuciform, which are often of considerable diagnostic value in the study of profiles. (See p. 249.)

Swelling takes place when the moisture content of dry soil is increased. The swelling curve, however, is not coincident with the shrinking curve. At the same moisture content the volume shown on the swelling curve is greater. D. P. Sideri\textsuperscript{15} distinguishes between (a) capillary or intermicellar imbibition in which water enters the spaces between the micellar structures and (b) intramicellar swelling, in which the actual micelles increase in volume.

PLASTICITY

Plasticity may be defined as the ability of a material to change shape continuously under the influence of an applied stress, and to retain the impressed shape on the removal of the stress. The exhibition of this property by soils is dependent on the presence of certain proportions of water. A. Atterberg\textsuperscript{16} has distinguished an upper and a lower plastic limit. The upper limit of plasticity of a soil is the point at which the soil-water paste is just fluid, i.e., the point at which it fails to retain the shape impressed on it by an applied force. The lower limit of plasticity is the point at which cylindrical portions of a soil-water paste crumble down on being rolled, i.e., the point at which the paste is unable to change shape continuously under the influence of an applied force. The difference between the moisture contents, calculated on the dry soil, at the upper and lower limits is termed the plasticity number. Since soils containing moisture in excess of the lower limit are plastic, this point represents the highest moisture content at which a soil can be cultivated without injury to tilth.

In Table XVIII, the plasticity data obtained by Atterberg for some typical Swedish soils are given.
Table XVIII. — Plasticity Data for Some Typical Soils (Atterberg)

<table>
<thead>
<tr>
<th>Soil</th>
<th>Upper Plastic Limit % Water</th>
<th>Lower Plastic Limit % Water</th>
<th>Plasticity Number</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Highly Plastic Soils</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Silurian clay</td>
<td>...</td>
<td>67</td>
<td>40</td>
</tr>
<tr>
<td>Ancylus clay</td>
<td>...</td>
<td>57</td>
<td>30</td>
</tr>
<tr>
<td>Glacial clay</td>
<td>...</td>
<td>51</td>
<td>26</td>
</tr>
<tr>
<td><strong>Moderately Plastic Soils</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fresh-water clay</td>
<td>...</td>
<td>52</td>
<td>37</td>
</tr>
<tr>
<td>Arable soil</td>
<td>...</td>
<td>42</td>
<td>30</td>
</tr>
<tr>
<td>Glacial clay</td>
<td>...</td>
<td>32</td>
<td>25</td>
</tr>
<tr>
<td><strong>Feebly Plastic Soils</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Arable soil</td>
<td>...</td>
<td>64</td>
<td>58</td>
</tr>
<tr>
<td></td>
<td>&quot;</td>
<td>33</td>
<td>28</td>
</tr>
<tr>
<td></td>
<td>&quot;</td>
<td>22</td>
<td>18</td>
</tr>
</tbody>
</table>

Plasticity is associated with the presence, in soils and clays, of inorganic colloidal material. The property can be exhibited also by finely ground lamellar minerals such as mica and biotite. It is more strongly developed in soils with siliceous clay fractions than in soils whose clay fractions contain notable proportions of sesquioxides.

COHESION

Cohesion is closely associated with plasticity, for plastic soils are cohesive and set into hard clods on drying. It is markedly affected by changes in moisture content and, in heavy and medium-textured soils, increases as the soil dries. The intractable clods formed in the drying out of heavy clay soils are familiar to cultivators. With extremely light sands, however, cohesion may pass through a maximum with decreasing water content and then decrease as complete dryness is approached. This is exemplified by the looseness of sandy soils when completely dry.
S. Johansson, using a method devised by Atterberg for the measurement of cohesion, has constructed curves connecting cohesion with moisture content for typical soils. The curves appear to be composite in character, but the work of W. B. Haines throws some doubt on the discontinuous character of the cohesion-moisture relationship.

SOIL STRUCTURE

The ability of soils to assume a condition of aggregation, thereby acquiring a definite structure, has been already mentioned. We shall now discuss the problem with more close attention.

In the first place we may distinguish the single-grain from the aggregated, granular, or crumb structure, and it is desirable to direct attention at once to a possibility of confusion. Whilst structure is generally regarded as a consequence of the aggregation of the primary particles of the soil into compound aggregates, there is a series of formations, also termed structures, in which the soil is predominantly in the single-grain structure. Such are the laminated and pulverulent structures shown by A horizons of podsols, and the columnar and prismatic structures shown by the B horizons of a group of soils termed solonetz (see Chapter XIII). Examples of these structures are shown in Plate III.

Columnar and prismatic structures are actually the result of the shrinkage consequent on the drying out of highly dispersed clays, and it is of interest to note that the structural units thus formed appear to be permanent over recurring cycles of moistening and desiccation.

Most of the investigations on structure that have been made during recent years, however, are concerned with structure in its more usual significance, namely, the aggregation of the soil into compound granular particles. Whilst it is generally recognized that the maintenance of the granular or crumb structure is desirable from the point of view of plant growth, considerable difficulty has been experienced in
explaining the mechanism of crumb formation and in finding means of expressing it quantitatively.

Growing plants have an effect on soil structure partly through their residues, including organic matter sloughed off from growing roots, and partly as a direct result of root action. This may be seen in the soil in which seedlings have been raised in close array. I. P. Sarakhov\textsuperscript{19} considers that grasses favour stable structure, perennial ryegrass being the most effective in this respect. Lucerne is the best structure-former among the legumes. Maize is considered to have a bad effect on structure. The good effect of grass on structure is familiar to every farmer who has ploughed up old turf. After years in grass, even the heaviest clay soils acquire a good structure.

The microbial population of the soil has an effect on structure formation. T. C. Peele\textsuperscript{20} attributes a binding action to the mucus produced by bacteria, e.g., those from pea nodules. The aggregates thus formed are water-stable. Addition of sucrose to the soil by stimulating microbial growth increases the percentage of water-stable aggregates. J. R. McHenry and M. B. Russell\textsuperscript{21} studied the effect of decomposition of untreated and hot-water extracted organic matter on aggregation. They found an initial increase in water-stable aggregates up to six days due to the decomposition products and microbial wastes. Then followed a breakdown and afterwards a further rise due to accumulation of decomposition products of the more resistant organic constituents. Investigations by J. P. Martin\textsuperscript{22} on a fungus of the \textit{Cladosporium} group showed a marked aggregation of silt and clay partly due to substances produced by the fungus and partly to the binding effect of the fungal mycelium. In the case of an aerobic bacillus of the \textit{B. subtilis-mesentericus} group, 80\% of the aggregating effect was due to substances exacted. In the case of the bacillus a hemicellulose-like substance was primarily responsible but being readily attacked by bacteria and actinomycetes, its aggregating effect is only fugitive in character. G. M. Browning
and F. M. Milan found that the aggregation caused by the decomposition of organic materials added to the soil reached a maximum effect in 20-30 days and gradually became ineffective.

With regard to the mechanism of crumb formation, it is generally recognized that the colloidal constituents play an important part in facilitating aggregation. In this connexion it is necessary to emphasize the distinction between the aggregates formed by humus and those formed by clay as a cementing agent. Where the aggregation is mainly due to humus the structure is loose and spongy, a good example being the structure found in many old garden soils and in potting-moulds. This structure, for which there is no satisfactory English term, is termed by the German farmer "Bodengare," the underlying idea being that the soil is "puffed up." Indeed, earlier theorists have actually attributed this structure to the evolution of carbon dioxide within the soil.

Where clay is a cementing agent, the structure is less spongy in character and the individual aggregates tend to be of a clod-like character, the "shot" and "nut" structures of the American workers being examples (see Plate III). In most agricultural soils both clay and humus participate in crumb formation.

Different types of aggregation have been recognized. A. T. Tiulin has distinguished true and false aggregates, the former held together by bivalent and tervalent cations and the latter by univalent cations or simple cohesion. A. Demolon and S. Hénin draw a similar distinction, but regard the looser type of aggregation (état molleux) as formed under the influence of compression. The loose aggregates can be considered as temporary in character, owing their existence to the operations of cultivation, the action of frost and possibly to the effects of plant roots. The other type of aggregates appears to be more permanent. We may regard it as due to the presence of colloidal material having exchangeable cations.
E. W. Russell\textsuperscript{26} has put forward the hypothesis that the cohesion, leading to crumb formation, exhibited at low moisture contents is due to the effect on clay particles of oriented molecules of water. The hypothesis applies equally to other polar liquids. In a dilute clay suspension each particle is surrounded by water molecules oriented along the lines of force of its electrostatic field. Free ions in the dispersion medium are similarly surrounded by oriented water molecules. If the suspension becomes concentrated

![Diagram](image)

**Fig. 13.—Orientation of water molecules between two clay particles. (Russell.)**

there will be an increasing tendency for water molecules to become oriented as shown in *Fig. 13*, where the negatively charged surfaces of two clay particles are shown linked through a pair of cations. The linkage system thus consists of: clay particle—oriented water molecules—cation—oriented water molecules—clay particle.

Crumb formation thus depends (1) on the presence of the double layer consisting of an inner shell of anions and an outer shell of cations, of which some are dissociated, and (2) on the presence of a polar dispersion medium. It cannot be expected with materials having no base exchange properties
nor by the desiccation of suspensions in non-polar liquids such as carbon tetrachloride.

Whilst it would appear to be exceedingly difficult to obtain a quantitative measure of the loose aggregation of the soil, apart from pore-space, which will be the greater, the more the soil is in this state, it might appear feasible to obtain a measure of the extent to which the soil is formed into true aggregates.

The specification and measurement of soil structure is a problem that presents many difficulties, for, whilst it is not difficult to be aware of the existence of structure, its precise expression is less easy. The solution of the problem has been attempted along a number of lines.

(1) There are many methods designed to express the aggregate analysis of soils, i.e., the relative proportions of the different sized crumbs present. Essentially these methods are similar to mechanical analysis without dispersion. The inherent difficulty in all the methods proposed is that in treating the soil, either by sieving or by sedimentation methods, there is an unavoidable interference with the natural structure, so that comparable results can be obtained only by a rigid prescription of the technique to be followed. A number of methods depend on dry sieving. There is also a variety of wet sieving methods, of which that of A. F. Tiulin\textsuperscript{24} is most widely used. The soil, after a preliminary wetting, is placed on a nest of sieves which are dipped 30 times in water. Whilst all the methods proposed have a certain value, they are essentially conventional, and until some general agreement is reached as to the precise details of the technique, the existing methods must be regarded as provisional. A. Demolon and S. Hénin\textsuperscript{25} reject the use of water alone in aggregate analysis and propose the use of 0.1\% calcium nitrate solution, by which, it is held, the original state of aggregation is maintained.

(2) Aggregate analysis, however carried out, does not give significant information unless the results of mechanical analysis with complete dispersion are available for compari-
son. The somewhat drastic measures used in this preliminary dispersion are evidence of the stability of the aggregates that are broken down. P. Vageler and F. Alten\textsuperscript{27} have proposed to express soil structure by a single factor, the \textit{structure factor}, which is simply the ratio of the difference between the total clay and the clay obtained by water dispersion expressed as a percentage of the total clay obtained by complete dispersion. With poor structural development will be a tendency for the ratio to approach zero. A measure of structure is also obtained where mechanical analysis is carried out with a simple water dispersion, the so-called aggregate analysis. The difficulty lies in prescribing the stage to which dispersion shall be carried. There is no evidence that the dispersion secured in the International B method, for example, represents more than an arbitrary stage in dispersion.

(3) The problem of soil structure may be approached from a different angle by studying the pore-space distribution. A simple expression of this is the division of the total pore-space into non-capillary pore-space (macro-porosity) and capillary pore-space (micro-porosity). A more exact expression may be given by an interpretation of the curves connecting moisture content with suction pressure. The form of the curve connecting $pF$ with moisture content can throw light on the distribution of pore-spaces in the soil. The relationship is complicated by hysteresis effects and by swelling and shrinkage of colloidal material with changing moisture content.

An important aspect of structure is the stability of aggregates. It is obvious that the loose or false aggregates have slight stability. D. G. Vilensky\textsuperscript{28} contrasts aqueous stability, the property whereby the soil disaggregates under the influence of water, and mechanical stability whereby resistance is opposed to disaggregation by mechanical means. It is suggested that both types of stability depend on the moisture content of the soil at the moment of formation of aggregates. This, of course, refers only to an individual
soil, for the stability of aggregates, compared among differ-
ent soils, depends on the nature of the colloidal complex
and the exchangeable cations. It is commonly supposed
that calcium-saturated colloids give the most stable aggre-
gates, an impression which is favoured by the known ameliorative effect of lime on the structure of soils. Recent
work has shown that base-unsaturation does not necessarily
imply bad structure. Thus J. F. Lutz\textsuperscript{29} considers that so
far as flocculation and low hydration are concerned, hydrogen
clays should possess better physical condition than calcium
clays. H. E. Meyers\textsuperscript{30} finds that calcium was no more effec-
tive than hydrogen in forming stable aggregates. The good
effect of lime is held to be due to its action on the organic
matter status of the soil.

**TYPES OF STRUCTURE**

Apart from the desirability of being able to give a quan-
titative expression of soil structure, it is very desirable for
many purposes, particularly for soil survey, to have methods
of describing soil structure as one element in soil mor-
phology. Many methods have been proposed, the majority
based on the work of S. A. Zakharov\textsuperscript{31}, whose three funda-
mental types are (1) cubicular, (2) prismatic, and (3) plate-
like. A development of Zakharov's system is that of the
American Soil Survey Association. Two main types are
distinguished, namely, (1) *simple structures*, in which the
particles are unaggregated and natural cleavage planes are
not visible, and (2) *compound structures*, in which aggrega-
tion with visible cleavage planes is evident. The simple
structures are subdivided into (a) simple-grained, (b) amorp-
hous, (c) massive, i.e., particles consolidated to a uniform
mass, (d) pulverulent, and (e) puddled. The compound
structures are divided into the three types cuboidal, pris-
matic, and platy, following Zakharov. The cuboidal types
are further subdivided into the granular and fragmental, the
former with poorly defined edges and surfaces and the latter
with well-defined edges and surfaces. Further subdivision
of these sub-types and of the columnar and platy types is on the basis of the size of the aggregates. A convenient distinction among the cuboidal types is between the rounded and the polyhedral aggregates. The rounded aggregates or crumbs may be distinguished by their size and resistance to crushing. Thus one may have coarse soft crumb, medium soft crumb, coarse hard crumb, etc. Hard crumb structure is generally termed granular, as in the chernozems.

In describing soil structure it is also necessary to note porosity (size and disposition of pores) and compaction. Attention must also be given to the presence of coatings, e.g., of humus or salts on the surfaces of the structural elements. W. Kubiena\(^3\) regards soils as built up of fabric units, which may be crystal fragments, raw humus particles, films, coagels, bridge or cavity-filling formations, etc. He distinguishes skeletal and plasmatic elements. The following types of aggregates are recognized:

*Porphyropeptic.* Uncoated minerals in an almost massive or at least only finely cracked ground mass which easily separates. This structure occurs in laterites.

*Porphyropeptic.* Minerals always covered by colloid coatings or thick films, with the appearance of being embedded in a ground mass. Characteristic of Mediterranean red earths.

*Intertextic.* Uncoated minerals bound by friable colloid bridges or in a ground mass with many cavities, as in the chernozems.

*Plectoamictic.* Mineral grains with colloidal coating bound by colloid bridges or in a cavernous ground mass.

*Chlamydomorphic.* Each grain with a colloid coating, bound together in loose aggregates, as in the B-horizon material of podzols.

*Agglomeratic.* Grains uncoated or with loosely-attached friable coatings. Plasmatic material present as aggregates in the pore-spaces. Aggregation undeveloped or only slightly developed.
"Bleicherdegefüge." Uncoated grains, as in bleached horizons of podsoils, with independent raw-humus agglomerates. No aggregate formation.

The structure of soil as viewed under field conditions, particularly in the examination of soil profiles, is an important aid to the diagnosis of soil characters. It should not, however, be given undue weight and regarded as other than an outward manifestation of the essential constitution and properties of the soil. And, therefore, whilst structure may be of the greatest utility for the diagnosis of soils in the field, it would be unwise to use it as an actual basis of classification. For example, soils with a certain type of structure have been termed "solonetz" by the Russian workers. At a later stage it was found that these soils were characterized by having sodium as the predominant cation in the absorbing complex and a hyperalkaline reaction. C. F. Shaw and W. P. Kelley have encountered examples in Utah and California, and J. H. Ellis and O. G. Caldwell examples in Alberta, of soils with solonetz type of structure, but with an entirely different type of cation complex. Which is the more significant, the chemical constitution or the visible structure? In the writer’s opinion, the constitution of the soil as revealed by chemical examination should undoubtedly have priority of importance.

COLOUR

The colour of the soil, especially the changes in colour through the horizons of the soil profile, is one of the most important aids to the recognition and description of the different genetic groups.

In the first place, it should be remarked that there is always a progressive lightening in colour as a soil dries out. The change may be attributed to the development of air spaces in the soil crums or in the colloidal coating of the coarse particles and, finally, to the partial dehydration of the colloidal material itself. Until a soil has dried down to the point at which air spaces occur there is, thus, no change
in colour. In laboratory experiments, the beginning of
colour change is found to correspond with the point at which
there is a change in the direction of the curve connecting
volume with moisture content. When the soil has reached
the stage of desiccation at which only hygroscopic moisture
is present there is no further change in colour.

Broadly speaking, it may be said that the most marked
differences between wet and dry soil colours are found in
soils containing high proportions of colloidal material. For
example, black peaty soils may become light grey when dry.

The colours of moist soils are generally more vivid and
strongly contrasted than those of dry soils. Soils which stand
out sharply from each other under moist field conditions are
much less distinct when they become dry laboratory samples.
It is therefore preferable to use the colour of the moist soil
for descriptive purposes where possible. The drying to
which soils are subjected in the preparation of laboratory
samples appears to involve colour changes which are not
readily reversible on re-moistening.

The colour of the moist soil is governed principally by
two constituents, namely, the colloidal organic matter and
the colloidal clay. Generally speaking, organic matter
confers the dark colour on soils. Among soils of the same
group, the colour becomes darker with increasing propor-
tions of this constituent. It should be noted that the effect
of organic matter is more marked in sandy than in silty or
clay soils. A coarse sand with 10% of organic matter may
be nearly black when moist, whilst the same proportion of
organic matter in a clay soil would have only a slight effect
on the mineral colour.

The darkening effect of organic matter varies among the
great soil groups. In the chernozems, the rendzinas, and
certain tropical and sub-tropical soils with high lime-status,
the humification of plant residues appears to produce a very
dark-coloured humus. On the other hand, in the red soils
of the humid tropics, the humus appears to be lighter in
colour. It may be, however, that its colour is masked by
the highly coloured clay colloids of such soils. In the bleached horizon of podsol soils the organic matter would appear to be only slightly coloured. That differences exist in the colour of humus is only to be expected from the diversity of the processes whereby it is produced.

Except in highly organic soils, the colour is governed mainly by the soil mineral matter and, in particular, by the colloidal clay. In studying soil colour, it is generally convenient to destroy the humus by boiling with sufficient 6% hydrogen peroxide, whereby the true mineral colour may be observed. The removal of even small proportions of organic matter by this treatment results in an appreciable "brightening" of colour, whilst in the case of dark-coloured organic soils the colour change is frequently remarkable.

Considering the mineral colour of soils, we may say at the outset that the unweathered minerals of the soil contribute but slightly to soil colour, which is conferred usually by the inorganic weathering-complex. On the other hand, H. E. Annett has shown that finely divided magnetite is largely responsible for the dark colour of certain black cotton soils in India. Blue and green colours in subsoils may be attributed to the presence of ferrous compounds such as pyrites, marcasite, vivianite, and glauconite.

The mineral colour is mainly due to compounds of iron, and to a less extent, of manganese. Yellow, brown, and red colours are due to the presence of ferric oxides with varying degrees of hydration. Broadly speaking, red colours are associated with lower degree of hydration, such as are commonly encountered in tropical soils. Some effect on colour must, however, be attributed to the stage of aggregation of the hydrate of ferric oxide present. The effect of a coloured constituent is more evident in sandy soils than in soils of heavier texture. On account of the low specific surfaces of such soils, comparatively small proportions of hydrated ferric oxide are sufficient to confer a marked red or brown colour. The strong colours of certain soils of the Trias and Old Red
Sandstone formations and of certain desert soils do not therefore necessarily denote high proportions of ferric oxide. The most vivid red colours are found in soils of the humid tropics. Red colours may also occur in desert soils. High temperatures are apparently necessary for their development, which may be due to decomposition of the clay complex.

Soil colour is markedly influenced by the constitution of the clay complex. Generally speaking, soils in which the weathering-complex has originated in the presence of notable proportions of calcium compounds, such as the chernozems, rendzinas, chestnut earths, and grey desert soils, show, so far as the mineral portion of the soil is concerned, grey or greyish-brown colours. The clay complex of such soils is rich in silica, and it may be presumed that sesquioxides exist only in combination with silica and not in the free state.

Grey colours occur in the A horizons of podsol profiles, due to the removal of ferric oxide by acid percolating waters. The grey colour is often only evident after removal of humus by peroxide treatment.

Grey colours are found also in the A horizons of solöti or degraded alkali soils. In such soils, the leaching out of the sodium soil in the absence of calcium carbonate leads to a decomposition of the clay complex and an eluviation of sesquioxides, leaving an A horizon rich in residual silicic acid.

There is also a large group of soils in which, although notable proportions of iron are present, grey colours prevail owing to reduction to ferrous compounds. These are the soils of marshes and bottoms where free aeration is prevented by excess of moisture. Such soils occur not only in cool and temperate regions, but also in the tropics, where the grey mineral colour is frequently masked by dark organic matter.

In considering the colour of soils it is important to decide whether the colour reflects recent pedogenic process or
whether it is simply a rock colour. The red colour of the soils of the English Trias and Old Red Sandstone must be attributed to the parent rock and not to the operation of contemporary processes. It represents rather the result of the weathering processes of Triassic and Devonian times.

SPECIFICATION OF SOIL COLOUR

The accurate specification of the colour of soil is a matter of considerable difficulty both on account of the great subjective errors to which an estimate of colour is liable and also on account of the lack of any generally recognized scale of colours.

Numerous attempts have been made to arrive at a system of recording soil colours. The soil colours may be matched with a colour of known specification in a colour atlas such as those of Ridgway or Ostwald; or, alternatively, they may be matched with composite colours, obtained by the combination, in varying proportions, of standard colours. The method employed by J. G. Hutton may serve as an example.

The desired soil colour was obtained by combination, in a rapidly rotating disc, of varying proportions of four standard colours, namely,

Neutral 9 ... ... ... ... White
Neutral 1 ... ... ... ... Black
Red 4/9 ... ... ... ... Red
Yellow 8/8 ... ... ... ... Yellow

The colours were obtained from the Munsell Colour Co., of Baltimore, Md., and were accurately specified by the U.S. Bureau of Standards.

A selection from Hutton’s data for the colour value of some typical soils is shown in Table XIX.
### Table XIX.—Colour Value of Typical Soils (Hutton)

<table>
<thead>
<tr>
<th>Soil Colour Name</th>
<th>White %</th>
<th>Black %</th>
<th>Yellow %</th>
<th>Red %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grey</td>
<td>22.0</td>
<td>60.0</td>
<td>11.0</td>
<td>7.0</td>
</tr>
<tr>
<td>Dark grey</td>
<td>16.0</td>
<td>67.0</td>
<td>9.0</td>
<td>8.0</td>
</tr>
<tr>
<td>Yellow</td>
<td>11.5</td>
<td>19.0</td>
<td>45.5</td>
<td>24.0</td>
</tr>
<tr>
<td>Greyish-brown</td>
<td>14.0</td>
<td>50.0</td>
<td>22.0</td>
<td>14.0</td>
</tr>
<tr>
<td>Reddish-brown</td>
<td>7.0</td>
<td>44.0</td>
<td>18.5</td>
<td>30.5</td>
</tr>
<tr>
<td>Dark reddish-brown</td>
<td>3.0</td>
<td>72.5</td>
<td>10.5</td>
<td>14.0</td>
</tr>
<tr>
<td>Light brown</td>
<td>10.0</td>
<td>59.0</td>
<td>18.0</td>
<td>13.0</td>
</tr>
<tr>
<td>Dark brown</td>
<td>4.0</td>
<td>73.0</td>
<td>12.0</td>
<td>11.0</td>
</tr>
<tr>
<td>Dark greyish-brown</td>
<td>9.0</td>
<td>73.0</td>
<td>11.0</td>
<td>7.0</td>
</tr>
<tr>
<td>Red</td>
<td>3.0</td>
<td>35.0</td>
<td>20.0</td>
<td>42.0</td>
</tr>
<tr>
<td>Dark red</td>
<td>0.0</td>
<td>70.0</td>
<td>7.0</td>
<td>23.0</td>
</tr>
</tbody>
</table>

### Bibliography


CHAPTER IX
WATER RELATIONSHIPS OF SOILS

RETENTION OF MOISTURE BY SOILS

Soil, on account of its granular and colloidal character, is able to retain moisture received through rainfall or irrigation. This property is of the highest significance both in the control of the pedogenic processes and in the life of plants. The study of the moisture relationships of soils may therefore be considered to be the most important group of problems in soil physics.

These problems fall under the following headings:—
(1). The manner in which water is retained by soils and the quantitative expression of the differences in their behaviour in this respect.
(2). The movements of water in soils, including the distribution within the soil of water received by rainfall or irrigation, and the factors affecting losses by evaporation.

In the earlier literature the retention of water by soils was viewed as the adherence of water films to the soil particles or, alternatively, as comparable with the retention of water in capillary tubes of varying cross section. This approach proved unsatisfactory, and in order to show more precisely the nature of the physical forces involved, theoretical and experimental studies have been made of water retention and movement in the ideal soil,* which was imagined as consisting of uniform spheres in close (hexagonal) packing.

The ideal soil concept was largely developed by W. B. Haines¹, who showed that the interstitial spaces in a system

*The epithet is used in a purely philosophical sense, and does not imply excellence for the growth of plants.
of close-packed uniform spheres were of two types, namely, tetrahedral and rhomboidal, the former occurring at the centre of four adjacent spheres arranged as an equilateral pyramid and the latter occurring between a group of six adjacent spheres arranged in two parallel planes. The spaces are present in the proportion of two tetrahedral cells to each rhomboidal cell. At low moisture contents water is present as rings at the points of contact of adjacent spheres. With increasing moisture the rings become larger until their edges come into contact. At this stage, it is shown that 24% of the pore-space is occupied by water. Up to this stage the water is described as pendular water. With further increase in moisture the liquid phase now forms a continuous system. The air phase is similarly continuous. This is described as the beginning of the funicular stage. With further increase in moisture the pores become full of water, and this is described as the capillary stage. It is shown that these transitions are not continuous but that discontinuous redistributions occur.

When the moisture decreases from saturation similar discontinuous changes in moisture distribution occur until the pendular stage is reached. Both in moistening and drying the moisture content is controlled by pressure deficiency, but different moisture contents can be associated with a given pressure deficiency, depending on the previous course of events.

In actual soils the shape and distribution of interstitial spaces will differ considerably from that in an ideal soil of uniform closely packed spheres, but the general relationships between moisture content and pressure deficiency with increasing and decreasing moisture content are not dissimilar.

Whilst conditions approximating to those in the "ideal soil" can scarcely ever be realized in actual soils, the treatment of soil moisture from this standpoint marks a considerable advance on the older theories and has thrown considerable light on the distribution and movement of water in actual soil. A discussion of this theory is outside the
scope of the present book and the student may be referred to
the discussions of W. B. Haines and R. A. Fisher. In natural soils, the water relationships are dominated
by the presence of colloidal material. The moisture asso-
ciated with soil colloidal material has been termed "im-
bibitional moisture" or "water of imbibition." The
imbibitional moisture has been considered to consist of gel
water associated with the colloidal material itself and "vesi-
cular water" held in the interstices of the gel structure.
The work of Hofmann, Endell, and Wilm (p. 133) has
shown that water is actually stored between the cleavage
planes of the clay mineral montmorillonite, which can alter
their spacing according to the amount of moisture present.
S. Mattson regards imbibition as essentially an osmotic
phenomenon depending on the presence of active exchange-
able cations and has propounded a theory of water retention
based on the Donnan equilibrium. It is now generally held
that "imbibitional moisture" cannot be considered as a
definite category.

It is evident that soils differ considerably in their
capacity for water retention. Thus, a heavy clay soil with
10% of moisture might be apparently dry, whilst a light
sandy soil with the same amount would be visibly moist.
Different stages of moistening have been distinguished and
the capacity of soils for moisture retention has been com-
pared by the use of defined states of moistening, varying
from the moisture retained by air-dry soils to the maximum
water-holding capacity of a soil in immediate contact with a
free water surface. In the ensuing sections we shall discuss
some of the measures proposed for characterizing the capacity
of soils to retain moisture.

It should be remarked at the outset that there is some
choice of methods of expressing the water content of a soil.
If the weight basis be chosen, the result may be expressed
either on the moist soil or on the dry soil. For most pur-
poses, the latter mode of expression is preferable, since it
gives a truer picture of the amount of water associated with a
unit weight of soil. For some purposes, it is convenient to express results as a volume percentage. This method is appropriate when results are to be considered in connexion with pore-space. Here the results may be calculated to the volume of the total soil or to the volume of the actual soil particles. The relationship of these modes of expressing water content may be seen from the following example (Table XX) for a hypothetical soil having apparent density 1.25, true density 2.50, and pore space 50%.

**Table XX. — Corresponding Percentages of Water by Weight and Volume for a Typical Soil**

<table>
<thead>
<tr>
<th>Percentage moisture by weight</th>
<th>Percentage moisture by volume</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moist soil</td>
<td>Dry soil</td>
</tr>
<tr>
<td>28.6</td>
<td>40</td>
</tr>
<tr>
<td>24.4</td>
<td>32</td>
</tr>
<tr>
<td>19.4</td>
<td>24</td>
</tr>
<tr>
<td>13.8</td>
<td>16</td>
</tr>
<tr>
<td>7.4</td>
<td>8</td>
</tr>
</tbody>
</table>

**Hygroscopic Moisture**

When a sample of soil, dried in an ordinary atmosphere, is placed in an oven at 100-105°C., or in a completely dry atmosphere, a small amount of water is driven off. This is called *hygroscopic moisture*, and is believed to be present in the form of thin films of approximately molecular thickness. The percentage of hygroscopic moisture in a given soil varies with the state of saturation of the atmosphere with respect to water vapour. In equilibrium with a perfectly dry atmosphere, the hygroscopic moisture is completely lost, but as the degree of saturation or relative humidity of the atmosphere with which the soil is in equilibrium increases, the moisture content of the soil also increases. This relationship has been investigated by M. D. Thomas⁴ and by A. N. Puri, E. M. Crowther, and B. A. Keen⁵. The curves
obtained are of a general sigmoid type with a point of inflexion corresponding with 50% humidity. Many of the earlier investigators attempted to determine the moisture content of the soil in equilibrium with a saturated atmosphere, believing that the figures thus obtained represented the hygroscopic capacity, i.e., the maximum amount of moisture that the soil could retain by hygroscopicity. The work of the above-mentioned authors shows that this conception is erroneous. Even if the experimental difficulties of maintaining an atmosphere in a state of saturation at a constant temperature were surmounted, the soil exposed to such an atmosphere would continue to take up moisture until it (the soil) became completely saturated. The curves connecting moisture content with relative humidity are in fact composite in character, and the point of inflexion may represent approximately the change from the curve indicating hygroscopic to that indicating capillary condensation.

H. A. Wadsworth⁶, in a discussion of the curves connecting moisture content with surface force, considers that the sorption in the lower part of the moisture relative humidity curve may be considered as relating to water held by "chemosorption," i.e., by surface adsorption, by hydration of ions on active surfaces, by sorption on crystal surfaces of minerals, including planar water, and also a certain proportion of the moisture held by organic matter. In the upper part of the curve moisture is considered to be held in the fine pores of the soil material and increases rapidly as relative humidity increases.

The degree of saturation of a closed atmosphere can be controlled by using equilibrium solutions, generally sulphuric acid, of varying strength. Thus, in the Bureau of Soils, Washington, 99.0% saturation is obtained by using 2% (by vol.) sulphuric acid. Puri, Crowther, and Keen propose 50% saturation, attained in equilibrium with sulphuric acid of density 1.3321. It is shown that the value obtained in equilibrium with a 50% saturated atmosphere does not differ greatly from the moisture content of ordinary
air-dry soil, since there is a fairly wide range on either side of this point over which the hygroscopic moisture content does not vary greatly. The significance of hygroscopicity determinations has been discussed by A. N. Puri.

The hygroscopic moisture content as ordinarily determined on air-dry samples, although it gives an approximate relative measure of a soil's hygroscopicity, is influenced by changes in atmospheric humidity. The same sample of soil may give different values on different days if there is any marked change in atmospheric conditions. Yet, on considering a large number of results, it is evident that the highest figures are given by soils rich in clay or organic matter and by soils rich in sesquioxides, such as laterites.

B. A. Keen and J. R. H. Coutts examined the correlation of the moisture content of soils in equilibrium with an atmosphere of 50% humidity with other soil constants. Whilst a high correlation was found between hygroscopicity and clay content, the correlation between hygroscopicity and ignition loss was less close. Since ignition loss is, except in heavy clay soils, mainly due to organic matter, this result might appear to imply that organic matter makes a small contribution to the hygroscopic capacity of the soil. Against this may be set the high values actually obtained for the hygroscopic moisture of peat soils. In some further studies on Natal soils, J. R. H. Coutts found a much higher correlation between moisture content at 50% humidity and ignition loss, but this was not surprising in view of the clay character of the soils used, for the loss on ignition in such cases is mainly due to water of constitution. Further work is desirable to establish the correlation which undoubtedly exists between hygroscopicity and organic matter content. The data obtained hitherto do not permit the accurate expression of hygroscopicity as a function of clay and organic matter content, but merely indicate correlations.

In a soil low in colloidal material, hygroscopic moisture is retained as films of approximately molecular thickness on the surface of the mineral particles, and the amount present
under standard conditions is a measure of the total surface. In ordinary soils, however, there is a certain amount of colloidal material and this is principally responsible for hygroscopic moisture retention.

E. A. Mitscherlich\textsuperscript{11}, who has used hygroscopicity as a measure of the specific surface, has attempted to distinguish between the total surface, which is concerned with the absorption of hygroscopic water, and is mainly a property of colloidal material, and the exterior surface of the mineral grains and soil crumbs. He has sought to obtain a measure for exterior surface by determining the amount of absorption from an atmosphere containing an organic vapour such as benzene, it being assumed that the benzene does not enter the interior micellar structures of the colloid material. R. K. Schofield\textsuperscript{2} considers, however, that the difference in the behaviour of the soil to water and to liquids such as benzene is one of degree rather than of kind. The theoretical basis of Mitscherlich's method and of other methods depending on similar comparisons is thus open to question.

An air-dry soil is incapable of yielding water to plants or of permitting microbiological activity. Further, although certain changes have been observed after long storage of air-dry soils, the rate of chemical change is inappreciable compared with the rate in moist soils, i.e., soils which contain moisture in excess of that retained by hygroscopicity.

**MOISTURE CAPACITY**

From the standpoint both of soil genesis and also of plant nutrition, the investigator is principally interested in moist soils. It is only in such soils that the pedogenic processes and the complex changes, involving the microbiological population and surface vegetation, take place.

It is a matter of common observation that even under similar climatic conditions, soils vary greatly in their water relationships. Generally speaking, sandy soils are dry and readily allow the rainfall to drain away from them, whilst clay soils are wet and retentive of the moisture obtained
from the rainfall. It is important to understand the mechanism of water retention, and also to possess methods of comparing this property in different soils.

A measure of the water-holding capacity of soils once widely used is the so-called maximum moisture capacity, obtained by determining the amount of water which a given quantity of soil can retain when placed in a shallow cylindrical vessel with a perforated bottom, saturated with water, and allowed to drain freely. The figure obtained represents the amount of water which a soil can retain in equilibrium with and in the vicinity of a free water surface. Various modifications of Hilgard’s original method have been proposed, for example, that of B. A. Keen and H. Raczkowski, but it should be noted that the results obtained correspond closely with the pore-space, for under such conditions on the larger voids will be empty of water. Further, the determinations are made on soil that is not in its field structure.

**MOISTURE EQUIVALENT**

When a soil is saturated up to its maximum water-holding capacity a large proportion of the water is only loosely held in the capillary spaces. In the determination of moisture equivalent, soil is placed in cups with perforated bottoms and submitted to a centrifugal force equivalent to 1000–2000 times that of gravity. The water held in the coarser interstices is thus removed and the only moisture retained is that in the smaller interstices associated with the soil colloids.

G. J. Bouyoucos has proposed a simplified method for determining moisture equivalent by submitting wet soil to suction on a Buchner funnel and determining the moisture content after water has ceased to drain away. Precautions are necessary to avoid errors by evaporation.

A. F. Lebedev has shown that the value obtained for the moisture equivalent varies with the centrifugal force used and proposes the use of 18,000 g., at which a limiting value,
which he terms the maximum molecular water capacity, is reached. It may be doubted whether the figure thus obtained represents a limiting value. Theoretically there should be a continuous decrease in moisture content with increase in the centrifugal force applied.

"STICKY POINT"

F. Hardy has proposed, as a measure of the water capacity of soils, the moisture content of a soil-water paste which can just be worked in the hands without being adhesive—the so-called sticky point. This is considered to represent the point at which the maximum imbibitional capacity of the soil colloidal material is satisfied. But since a sticky point can be obtained for fine sands and for kaolin, in which colloidal gel material is present in very small amounts, the moisture present at this limit must include some interstitial water.

B. A. Keen and J. R. H. Coutts have determined the sticky point of a number of soils from which most of the colloidal organic matter had been removed by treatment with hydrogen peroxide. The relationship with the clay content is approximately linear. Plotting the "sticky point" moisture, Sp, against the clay, the curve cuts the Sp axis at about 16% moisture. From this it may be deduced that a soil with no colloidal material should show a sticky point of approximately 16% moisture by weight. For an "ideal soil" with close packing the moisture content calculated on a weight basis with the interstices filled would be 14.6%.

B. de C. Marchand finds, for Cape Western Province soils, that the sticky point water, expressed as a volume percentage, is given by the equation

\[ Sv = 0.58 \cdot \text{Clay} + 26 \pm 1.2 \]

The value 26 in this equation is in complete agreement with the theoretical pore-space for an ideal soil—a result which can only be possible by reason of compensating adjustments, since the actual particles of soils are neither spherical nor of uniform size.
J. A. Prescott and H. G. Poole\textsuperscript{18}, working with Australian soils, found a tendency for the sticky point moisture to have a minimum value at 10% of clay.

The contribution of organic matter to the "sticky point" water capacity of soils has not yet been determined, but it is known that highly organic soils show high "sticky points."

It should be added that the moisture conditions in a soil at the "sticky point," as determined in the laboratory, are rarely, if ever, attained under natural conditions. In this determination, the moistened soil is brought into the single-grain or puddled condition and the paste consists simply of soil with water completely saturating the colloidal gels and filling the interstices.

**IMBIBITIONAL MOISTURE**

A. F. Joseph\textsuperscript{10} attempted to determine the imbibitional moisture in the clay fractions of a series of soils by obtaining for each soil the moisture equivalent and the xylene equivalent. Assuming that, in the case of xylene, there is no imbibition by gels, then the difference between the two values should give the imbibitional moisture, i.e., the moisture associated with the colloidal gel material.

Some results obtained by this method are given in Table XXIII.

**Table XXIII.—Moisture equivalent, xylene equivalent, and imbibitional moisture of clay fractions with different cations (Joseph)**

<table>
<thead>
<tr>
<th></th>
<th>Original Clay</th>
<th>Sodium Clay</th>
<th>Calcium Clay</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moisture equivalent by weight</td>
<td>68.7</td>
<td>125.0</td>
<td>63.0</td>
</tr>
<tr>
<td>Xylene equivalent by weight</td>
<td>11.7</td>
<td>11.4</td>
<td>11.1</td>
</tr>
<tr>
<td>Imbibitional moisture by volume</td>
<td>143.0</td>
<td>292.0</td>
<td>131.0</td>
</tr>
</tbody>
</table>

Some interesting points arise from these data. In the first place, it will be noticed that the xylene equivalent is
practically the same for the original clay, the sodium clay, and the calcium clay. The imbibitional moisture is, however, far higher in the sodium clay than in the other clays. This is in agreement with the highly dispersed character of sodium soils. The original soil, containing principally calcium as its exchangeable base, does not markedly differ from the calcium soil. The clay fractions of alkaline soils would probably approximate to sodium clays.

In Table XXIV are given a selection from Joseph's data for the imbibitional moisture of some representative clays. The correlation of the figures obtained for imbibitional moisture with the $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio may be noted. It should be added that the clay of china clay differs from the clay of ordinary soils. Soils having a clay $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio of 2.0 would retain appreciable amounts of imbibitional moisture.

**Table XXIV.—Imbibitional moisture and silica-alumina ratio for clay fractions (Joseph)**

<table>
<thead>
<tr>
<th>Clay Fraction</th>
<th>$\text{SiO}_2/\text{Al}_2\text{O}_3$</th>
<th>Imbibitional moisture Volume %</th>
</tr>
</thead>
<tbody>
<tr>
<td>China clay</td>
<td>...</td>
<td>2.11</td>
</tr>
<tr>
<td>Ball clay</td>
<td>...</td>
<td>2.66</td>
</tr>
<tr>
<td>Subsoil clay, N. Wales</td>
<td>...</td>
<td>2.84</td>
</tr>
<tr>
<td>Kassala alluvium</td>
<td>...</td>
<td>3.77</td>
</tr>
<tr>
<td>Badob</td>
<td>...</td>
<td>3.90</td>
</tr>
<tr>
<td>Blue Nile alluvium</td>
<td>...</td>
<td>4.53</td>
</tr>
<tr>
<td>Bentonite</td>
<td>...</td>
<td>6.96</td>
</tr>
</tbody>
</table>

**Categories of soil moisture**

From the foregoing account it may be seen that a considerable body of investigation has been directed towards the classification of soil moisture into definite categories. Thus, L. J. Briggs\(^{20}\) in 1897 proposed that soil moisture should be divided into three categories, namely, (1) hygro-
scopic moisture; (2) capillary water, i.e., water held by surface tension in films and in capillary spaces; and (3) gravitational water, i.e., water free to move downwards under the influence of gravity. Briggs and his co-workers developed this classification further with special attention to the capillary water, which was further subdivided according to its availability to plants. The idea of the *lento-capillary point*, the point at which the movement of soil moisture becomes so slow as to be ineffective was introduced.

G. J. Bouyoucos proposed a classification based on the idea of "unfree water," i.e., water held so closely by the soil colloids that it is not readily available to plants and freezes only at temperatures considerably below 0°C. This classification is briefly as follows:

**Gravitational water.**
Free water, freezing at not less than −1.4°C. and available to plants.

**Unfree water.**
   a. Capillary—freezes at −4°C., strongly held and slightly available to plants.
   b. Combined—does not freeze at −78°C., and is non-available to plants.

Bouyoucos's classification has been strongly criticized but served a useful purpose in directing attention to the great force with which the water of fine capillaries is held by the soil.

A. F. Lebedev proposed the recognition of the following forms of soil moisture:

- Water vapour.
- Hygroscopic water.
- Film water.
- Gravitational water.

This and similar classifications of soil moisture are all based on the capillary tube approach to the problem of soil moisture. It is now generally held that such classifications are arbitrary, and do not correspond with any real classification of soil moisture. We have already seen how
the concept of a definite hygroscopic capacity proved illusory. Similarly with such categories as imbibitional moisture, capillary water, and gravitational water, their definition can be made only in terms of conventional methods of determination. The modern tendency is to seek for the expression of the moisture status of soils in the form of continuous functions.

CAPILLARY POTENTIAL AND pF VALUE

The various measures proposed and used for characterizing water-holding capacity of soils all measure the wetness of soils under certain specified conditions. E. Buckingham first attempted to give this idea precision by introducing the term capillary potential on the analogy of temperature and electrical potential. Just as temperature differences determine the flow of heat, and electrical potential the flow of electricity, so capillary potential may be supposed to determine the translocation of moisture in soils. Thus, if at two adjacent points in the soil there is a difference of capillary potential, moisture movement should result. Buckingham's capillary potential was expressed in terms of the equivalent water column corresponding with the suction pressure or pressure-deficiency required to cause water movement from the soil, or the positive pressure for water movement into the soil at a given moisture content. Capillary potential is a measure of the difference in free energy between water in bulk and soil water under the given condition. It may be defined as the work required to move a unit mass of water against capillarity from the level of a free water surface, i.e., the water-table, to a given point above this level.

It is necessary to point out that inasmuch as capillary potential increases with decreasing moisture content, water movements will tend to take place from points of low to points of high capillary potential. Capillary potential is thus an inverse measure of wetness and the analogy with temperature and electrical potential is actually reverse. It is important to bear this in mind in the following discussion.
Capillary potential is conveniently expressed as suction pressure in centimetres of water. The relationship between suction pressure and moisture content in the range, corresponding with suction pressures up to about 100 cm. of water can be easily studied by means of a simple apparatus due to W. B. Haines\(^{24}\). The soil is placed in a layer of about 2 cm. on a Buchner funnel, the bottom of which is covered by a sealed-on filter paper. The funnel is connected with a burette-manometer as shown in Fig. 14. The apparatus is filled with water and, by varying the height of the burette, varying pressure deficiencies, indicated by the difference between the height of the water in the burette and the mean height of the soil layer, can be obtained. The burette reading serves to indicate withdrawal of water from the soil as suction pressure is increased. At first very little water is withdrawn from the soil, corresponding with the stage when concave menisci are formed at the surface of the
soil. In the next stage there is a considerable withdrawal of water corresponding with the evacuation of the larger voids. As successively smaller voids become empty, the rate of removal of water with increasing suction pressure slows down. The form of the curve connecting suction pressure with moisture content is characteristic and differs for different soils. If instead of plotting moisture content against suction pressure, the percentage void space is plotted, a curve is obtained giving the distribution of pore-spaces of different sizes.

In Haines’s method and others of the same type the suction pressure is given by the difference in height between the layer of soil and the level of the free surface in the other limb of a U tube. The situation is essentially similar in the case of a column of soil above a water table and in a state of equilibrium. Over the lower part of the range, the suction pressure increases proportionally with the height above the water-table and the moisture content falls off in correspondence. This, however, is true only over the range within which soil moisture movements are sufficiently rapid for equilibrium to be attained within the period of the experiment. At high suction pressures, i.e., above about 1,000 cms., movements are so slow that there is little change in moisture content with increasing height. The equilibrium is only apparent and, given sufficient time, adjustment would eventually take place.

R. K. Schofield\textsuperscript{25} has proposed the logarithm of Buckingham’s capillary potential as a measure of the moisture status of the soil and for this he proposes the term $pF$. It has the advantage of analogy with pH and enables the enormous differences in capillary potential between dry and moist soil to be compressed within a convenient range.

An example of the translation of a moisture constant into terms of $pF$ may be quoted from Schofield. In modification of a method proposed by G. J. Bouyoucos\textsuperscript{14}, he measures the amount of water retained after submitting wet soil to suction on a Buchner funnel so packed as to allow free passage of
water but not of air. Any desired pressure difference up to one atmosphere can be maintained across the filter. A suction pressure corresponding with one atmosphere is equivalent to that of a column of water about 1,000 cm. in height. The soil in equilibrium is said to have a pF of 3 ($\log_{10} 1,000$) and it is a simple matter, by making corresponding direct determinations of moisture content, to obtain a number of points connecting moisture content with pF values

**ROTHAMSTED GRASS PLOTS**

**(LOAM)**

- Vapour Pressure (Puri Crowther & Keen)
- Freezing Point (Schofield and Botelho da Costa)
- Direct Suction
- Centrifuge

**Fig. 15.**—Moisture-pF curve of Rothamsted soil. (Schofield.)
up to 3. To obtain higher pF values it would be necessary to use pressure above the filter. In the above experiment the starting point is from wet soil. By a variation in the method of experiment it is possible to determine the amount of moisture which the soil will absorb against a given pressure difference. It is found that different values are obtained for the moisture content at the same pF according to whether the soil is being dried out by suction or being wetted by
absorption against pressure. There are thus two curves connecting pF with moisture content, one for drying and the other for wetting. This is in harmony with the observed result that moisture movement does not necessarily result when there is a moisture gradient, or rather that movement may be so slow as to be unobservable within the period of an ordinary experiment. Schofield has used other data to obtain points on the pF-moisture content curve. Thus the depression of freezing point and the aqueous vapour pressure of soils can, by appropriate equations, be translated into terms of pF. These, together with values obtained by suction methods and by centrifugal methods, can be combined to give curves for wetting and drying. Such a curve is shown in Fig. 15, which relates to a Rothamsted soil, the upper curve indicating drying conditions and the lower wetting conditions. A similar curve deduced from published results is given in Fig. 16.

The discrepancy between the wetting and drying curves, usually referred to, but not explained, as an hysteresis effect, is to some extent due to the fact that the determinations were made on laboratory samples and not under field conditions. The drying curves were obtained with samples first air-dried and then flooded with water. The pore-space here is much greater than under field conditions. A. L. C. Davidson and R. K. Schofield²⁶, in an investigation of suction pressures with porous stones, found that equilibrium was more rapidly obtained under drying than under moistening conditions. The slow wetting of air-dry samples, however, gives conditions more closely approximating to those in the field.

The distribution of water in the soil by capillary transmission has been examined and formulated by E. C. Childs and H. H. Nicholson²⁷.

The conception of capillary potential translated into terms of pF, together with the recognition that for every soil there are two pF-moisture content curves, one for wetting and the other for drying, constitute an important clarification of our ideas on soil moisture problems. It is now
possible to have a convenient comparison of the moisture status of soils by means of pF values.

The implications of the pF conception have not been fully worked out, but one example may indicate its use. In the supply of water to a plant, transfer from soil to plant roots may be considered to result from a difference in capillary potential or pF. As the moisture content of the soil falls the capillary potential and pF increase. A point is eventually reached when the soil is no longer able to supply water rapidly enough to cover losses by transpiration and, with loss of turgor, the plant wilts. The point at which this occurs is termed the willling point. Schofield has shown, from a consideration of the data for moisture contents at the point of permanent wilting of plants, that this point may be represented by a pF of about 4.2 for a number of soils. Other critical moisture conditions can be similarly defined in terms of pF.

WATER MOVEMENTS IN SOILS

The earlier work on water movements in soils was dominated by the capillary tube hypothesis. Thus, W. H. Green and G. A. Ampt developed equations for the flow of water through soils based on Poiseuille's equation for the flow of liquids through capillary tubes, obtaining equations for downward percolation, upward motion, and horizontal motion. Experimental data with coarse materials were found to agree fairly well with the equations obtained, but complications were introduced by the erroneous assumption of complete saturation behind a moving water front, whilst still more serious difficulties were introduced in actual soils by the swelling of colloidal material on wetting. The same objections may be applied to the use of the ideal soil hypothesis to obtain equations for the movement of water in actual soils.

Probably the most fruitful conception in the study of water movements is that of capillary potential, or its more convenient form pF, introduced by Schofield. According to
this, water tends to move from a point of low to a point of high capillary potential. As has been seen, for each moisture content there are two pF values, one for wetting and the other for drying conditions. Thus a dry soil placed in contact with a wet soil will abstract water from it until its wetting pF is equal to the drying pF of the moister soil. A soil can thus be in capillary equilibrium with a soil of higher moisture content from which it is abstracting moisture.

It is important to realize that when water is added to a column of dry soil it will not diffuse throughout until an uniform moisture content is attained, but will only moisten a limited depth of soil. Beyond this limit there is a sharp decrease in moisture content. This was recognized by J. A. Widtsoe and W. W. McLaughlin in their proposed *lento-capillary point*, at which capillary movements become sluggish and ineffective.

Schofield pictures the absorption of water applied to a dry soil as involving two stages. In the first stage water runs rapidly through the larger interstices and is rapidly absorbed by the soil, which is moistened to a degree corresponding with a very low pF, and to a depth corresponding with the amount of water available. There is now a diffusion involving transfer from the wet upper layer to the dry lower layer. Movement into this layer where wetting conditions will obtain is against suction by the upper layer with drying conditions. After a time movement ceases, the limit being set by the pF value above which the rate of wetting of the dry soil is negligible.

**VARIATIONS IN MOISTURE STATUS IN THE SOIL PROFILE**

The capacity of a soil to retain water cannot be defined without reference to the soil horizons which lie below it. Just as, from the standpoint of soil genesis, the unit of study is the soil profile, so in studies of soil moisture, the system to be considered is the whole succession of horizons down to the ground-water level, or if this is very deep, down
to the limit of atmospheric influences. And, therefore, although laboratory studies on soil moisture problems are indispensable for the elucidation of the fundamental principles, the final synthesis can be made only face to face with actual soil profiles.

If a hole or trench be dug in the soil in a valley bottom, a level is reached at which there is complete saturation. If the hole is sunk still further, water will stand in it up to this level, which is called the water-table. The water-table usually follows the contours of the land in a less accentuated manner. The depth below the surface at which it is encountered increases as the sides of a valley or hill are ascended. In the bottoms of valleys it approaches the surface of rivers and lakes. Complications are produced by impervious strata, so that in some cases the water-table may be quite near the surface in elevated situations. This is sometimes called a "perched" water-table.

Owing to the predominance of rainfall over evaporation during the winter months, and of evaporation over rainfall during the summer months, the position of the water-table fluctuates, generally reaching a maximum at the end of the winter and a minimum at the end of the summer. Abnormal rainfalls or droughts may disturb this general succession.

In many soils, even in humid climates, the water-table lies so far below the surface as to have practically no effect on moisture conditions at the surface. Thus, in the chalk upland soils of England, the water-table is scarcely a factor in the soil moisture problem. It is difficult to assign a limit below which the water-table does not affect conditions in the surface soil. It would be fairly safe to assume that, where this level is more than twenty feet below the surface, the moisture relationships are mainly controlled by the conditions within the layer of free percolation. On the other hand, a water-table lying within four feet of the surface has a definite effect on the moisture relationships at the surface. For example, a certain coarse sandy soil in Anglesey is in agricultural use in low-lying situations where the water-table
is within three feet of the surface, whilst in slightly elevated situations it is blowing dune.

The influence of position relative to a water-table on water content is illustrated by data obtained by E. W. Hilgard in a sandy alluvium in California (Table XXII).

Table XXII.—Variation in moisture content in a sandy alluvium at different distances from the water-table (Hilgard)

(From "Soils," by permission of The Macmillan Co., Publishers.)

<table>
<thead>
<tr>
<th>Inches above water-table</th>
<th>1</th>
<th>6</th>
<th>12</th>
<th>18</th>
<th>24</th>
<th>30</th>
<th>36</th>
<th>42</th>
<th>47</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moisture %</td>
<td>36.6</td>
<td>35.0</td>
<td>32.5</td>
<td>27.6</td>
<td>21.4</td>
<td>15.3</td>
<td>12.0</td>
<td>10.3</td>
<td>4.3</td>
</tr>
</tbody>
</table>

The moisture content immediately above the water-table will, of course, correspond closely with complete saturation of the pore-space, as in the determination of maximum water-holding capacity, and the gradual decrease in moisture content towards the surface represents the changing equilibrium between gravity and capillary forces. The low moisture content at the surface may be attributed to loss by evaporation.

If we assume that the moisture content at each level represents equilibrium, any moisture in excess of this at any level is subject to gravity and tends to sink to the water-table. Some writers have drawn a distinction between capillary water and gravitational water, the latter being any water in excess of that which a soil can hold by capillary action at a particular level above the water-table.

We must now examine the distribution of soil moisture in a soil profile from the standpoint of what has been said above on the relationship between suction pressure and soil moisture. Let us consider a column of soil extending from the surface down to the level of the water-table. In ordinary moist soils with irregular space distribution, the moisture present forms a continuous system. Thus, except for a possible dry zone at the surface, all the moisture in the profile
is in capillary connexion. At the level of the water-table the pressure within the soil moisture is approximately that which would obtain at a corresponding free water surface. Above the water-table, however, the pressure falls off with height. Thus, at 1 cm. above the water-table there is a negative pressure or suction pressure equivalent to 1 cm. of water. For some distance above the water-table the suction pressure increases proportionally to the height, but this relationship holds for only a limited distance. The zone over which suction pressure is in linear relationship with height above water-table has been termed the “capillary fringe.” But here, as in other aspects of the problem of soil moisture, no hard and fast line can be drawn. In a perfectly uniform column of soil, the moisture content in the upper part of the profile, if in temporary equilibrium, would tend to correspond with a pF of about 3, the so-called field capacity, even although the water-table were more than 1,000 cm. below the surface. Near the water-table, or at the surface, where the water-table is at small depths, the equilibrium moisture content approximates more closely to the pF corresponding with the height above the level of complete saturation.

DISTRIBUTION OF ADDED WATER IN THE SOIL PROFILE

In very wet soils, a considerable proportion of the moisture present is free to move downwards under the influence of gravity and, eventually, an equilibrium distribution is attained depending on the position of the water-table and the moisture capacity of the different horizons.

The distribution of moisture applied to the surface of a dry soil is shown by some experiments by C. F. Shaw. Using 36 inches of dry soil and applying a 6-inch head of water, it was found that after 24 days the moisture content was fairly constant at about 20% down to 19 inches and then fell off to 3.33% at about 30 inches. After 123 days the moisture content decreased slightly from 17.32% at the surface to 16.16% at 24 inches. It was concluded that this represented practically the equilibrium distribution of the
water applied, apart from possible movements in the vapour phase. The average value, 16.79%, termed by Shaw the normal moisture capacity, approximates fairly closely to

![Percentage Moisture vs Depth in Inches](image)

**Fig. 17.**—Distribution of water in tubes receiving a 6-inch irrigation, after prolonged periods of adjustment. (C. F. Shaw.)

*From *Soil Science*—(by permission.)*

the moisture equivalent for the same soil, namely, 15.88%. These results are shown graphically in Fig. 17. It is suggested that, for moisture contents up to the normal moisture capacity, capillary movement is negligible. The normal moisture capacity of Shaw is thus apparently the lento-
capillary point of Widtsoe and McLaughlin, but appears to
lie above rather than below the moisture equivalent. The
subsequent drying out of the soil would take place entirely by
evaporation, which would not necessarily be entirely at the
surface, but might involve vapour movements from lower
levels. In this connexion, the influx and efflux of air conse-
quent on diurnal temperature variations is probably of
importance.

Shaw's results were obtained with a uniform soil under
laboratory conditions. In a natural profile, the variations in
the nature and content of colloidal material throughout the
profile would result in a different distribution of moisture.
Presumably the presence of a considerable proportion of
organic matter in the surface soil would be reflected in a
higher equilibrium moisture content. The march of the
curve down to the limit of wetting would, indeed, tend to
follow that of the moisture equivalent.

F. J. Veihmeyer and A. H. Hendrickson have studied
the field capacity of certain irrigated soils in California. By
field capacity is understood the "amount of water held in the
soil after excess gravitational water has drained away and the
rate of downward movement has materially decreased." For
all except the lightest soils, field capacity approximates
closely to the moisture equivalent as determined by the modi-
fication of the method of L. J. Briggs and H. L. Shantz. It
corresponds, according to Schofield, with a pF of about 3.
In the absence of evaporation, from two to three days after
irrigation are required for the field capacity to be reached,
but conditions are materially altered by the presence of im-
pervious layers. If an amount of water is applied either as
rain or by irrigation to a dry soil less than that to restore
it to field capacity throughout, when equilibrium is
reached, there is a fairly sharp transition from moistened to
unmoistened soil, all the moist soil showing approximately
the same degree of wetness relative to the moisture equiva-
 lent. This is readily intelligible when it is realized that
capillary transmission becomes negligible below a certain
moisture content, which may be, for practical purposes,
equated with Widtsoe and McLaughlin's herto-capillary point (p. 284).

In the presence of a water-table near the surface, the equilibrium moisture distribution is modified. In the Shaw experiment, a water-table at, say, 24 inches below the surface would be reflected in an increased moisture content, as in the Hilgard experiment mentioned on page 286. The presence of impervious or relatively impervious strata would have the same effect as a water-table.

**CAPILLARY RISE**

In the earlier literature on soil moisture, dominated by the capillary tube hypothesis, considerable importance was attached to the lift of water by capillary action from a water-table or a moist subsoil. It is now known that whilst such movements may take place to a limited extent in wet soils, in other words with soils at low pF values, a limit is soon set to the amount of rise which can thus take place.

Thus C. F. Shaw and A. Smith\(^3\) in an experiment with columns of Yolo loam of varying height above an artificially maintained water-table, found that with 4 foot tubes the average monthly loss was 3.595 cm., with 6 foot tubes 1.92 cm., with 8 foot tubes 1.0 cm.; and with 10 foot tubes only 0.165 cm. Thus with a water-table at 10 feet below, the capillary rise, which presumably controlled the amount of evaporation, was negligible. Similarly, B. A. Keen\(^4\) found that with 6 foot cylinders of soil sunk in the ground and originally saturated with water, the water-table in a heavy loam did not fall below 80 cm. In fine sand the limit was 70 cm., and in coarse sand 35 cm. from the surface. He concluded that capillary rise was ineffective over the depths indicated. An examination of the data in both of the above series in the light of the pF hypothesis would be of interest.

**LOSES OF MOISTURE BY EVAPORATION**

Under a covering of natural vegetation, losses from the soil to the atmosphere are principally by transpiration from the leaves of plants. Under crops, the losses are partly by
evaporation from bare soil. The rate of loss by transpiration and evaporation depends partly on meteorological factors such as temperature, humidity, and wind velocity, and partly on the character of the transpiring vegetation or evaporating soil. Other things being equal, moisture losses are greater under a covering of vegetation than with bare soil. The diminution in evaporation consequent on deforestation may lead to waterlogging. (See p. 362.)

The principal circumstance affecting the evaporation of moisture from soils is the temperature. This is illustrated by the data, reported by A. D. Hall, obtained from the Rothamsted lysimeters. The amount of percolation is measured from columns of soil in which the original structure has been preserved unaltered. The surface of the soil is maintained tilled but uncropped. The rainfall is completely accounted for by evaporation and percolation since there is no run-off from the surface. Hence, by measuring the percolation, the amount of evaporation can be determined. The data are shown in Table XXV.

<table>
<thead>
<tr>
<th>Month</th>
<th>Rainfall Average</th>
<th>Temperature 26 years average</th>
<th>Percolation through 60 inch gauge inches</th>
<th>Evaporation inches</th>
<th>Daily Evaporation inches</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan.</td>
<td>2.32</td>
<td>36.6</td>
<td>1.96</td>
<td>0.36</td>
<td>0.0116</td>
</tr>
<tr>
<td>Feb.</td>
<td>1.97</td>
<td>38.2</td>
<td>1.48</td>
<td>0.49</td>
<td>0.0175</td>
</tr>
<tr>
<td>Mar.</td>
<td>1.83</td>
<td>40.9</td>
<td>0.95</td>
<td>0.88</td>
<td>0.0284</td>
</tr>
<tr>
<td>Apr.</td>
<td>1.89</td>
<td>45.5</td>
<td>0.53</td>
<td>1.36</td>
<td>0.0453</td>
</tr>
<tr>
<td>May</td>
<td>2.11</td>
<td>51.2</td>
<td>0.50</td>
<td>1.61</td>
<td>0.0520</td>
</tr>
<tr>
<td>June</td>
<td>2.36</td>
<td>57.5</td>
<td>0.62</td>
<td>1.74</td>
<td>0.0580</td>
</tr>
<tr>
<td>July</td>
<td>2.73</td>
<td>60.7</td>
<td>0.65</td>
<td>2.08</td>
<td>0.0671</td>
</tr>
<tr>
<td>Aug.</td>
<td>2.67</td>
<td>59.9</td>
<td>0.58</td>
<td>2.09</td>
<td>0.0674</td>
</tr>
<tr>
<td>Sept.</td>
<td>2.52</td>
<td>55.9</td>
<td>0.76</td>
<td>1.76</td>
<td>0.0587</td>
</tr>
<tr>
<td>Oct.</td>
<td>3.20</td>
<td>48.0</td>
<td>1.68</td>
<td>1.52</td>
<td>0.0490</td>
</tr>
<tr>
<td>Nov.</td>
<td>2.86</td>
<td>42.6</td>
<td>2.04</td>
<td>0.82</td>
<td>0.0273</td>
</tr>
<tr>
<td>Dec.</td>
<td>2.52</td>
<td>37.7</td>
<td>2.04</td>
<td>0.48</td>
<td>0.0156</td>
</tr>
</tbody>
</table>
There apparently is a close correlation between temperature and the daily evaporation. R. J. Koshal\textsuperscript{38}, however, suggests that the high percolation in the winter months represents the accumulation of water during the wet autumn months, whilst the low figures in summer may be due to the delayed effect of the light spring rainfall. It cannot be doubted, however, that there is a considerable increase in the rate of evaporation with temperature. It can be readily understood that, in tropical climates, the rate is still greater. Where losses are due to transpiration from plants, the temperature effect will be similar.

Evaporation is markedly increased by wind, particularly if the atmospheric humidity is low. The rapid drying out of soils in east winds is well known to practical agriculturists.

The effect of variations in atmospheric humidity is shown by some data from Masure, cited by E. Ramann\textsuperscript{37} (Table XXVI).

**Table XXVI. — Effect of Varying Humidity on Rate of Evaporation from Soil (MASURE)**

<table>
<thead>
<tr>
<th>Temperature °C</th>
<th>Relative Humidity %</th>
<th>Evaporation mm.</th>
</tr>
</thead>
<tbody>
<tr>
<td>17.6</td>
<td>74</td>
<td>0.93</td>
</tr>
<tr>
<td>17.7</td>
<td>79</td>
<td>0.62</td>
</tr>
<tr>
<td>17.0</td>
<td>89</td>
<td>0.38</td>
</tr>
<tr>
<td>17.2</td>
<td>91</td>
<td>0.25</td>
</tr>
</tbody>
</table>

The great differences in the rate of evaporation depending on temperature, humidity, and wind velocity, involve corresponding variations in the amount of percolation through the soil. And therefore the rainfall, in itself, is a very inadequate measure of the leaching to which a soil is subjected. With a mean annual temperature of 5°C, varying between 0° and 10°C, a mean annual rainfall of 20 inches would be sufficient to furnish drainage. The same rainfall, with a mean annual temperature of 25°C, varying between 15°C and 30°C, would be insufficient to outweigh evaporation.
The principal soil factor affecting the rate of evaporation is the moisture content. Generally speaking, in laboratory experiments with thin layers of soil, the rate of evaporation is constant over a wide range of moisture content, but tends to fall off when the region of the hygroscopic co-efficient is approached. Under field conditions, the problem is more complex, for the rise of capillary water to the surface plays a part. But when the moisture content at the surface is reduced to the field capacity, further replacement from below becomes slowed down and subsequent losses include evaporation from the body of the soil down to the level at which capillary movement is still possible.

The successive effects of irrigation and evaporation are well shown by some observations of H. Greene in the
Sudan Gezira. The soil used was a heavy aeolian clay having the following soil moisture constants:

- Moisture equivalent about ... ... 40%
- Lento-capillary point about ... ... 28%
- Hygroscopic co-efficient about ... ... 14%

It will be noticed that the lento-capillary point for this soil lies considerably below the moisture equivalent.

The results of the investigation are shown diagrammatically in Fig. 14. Before irrigation, the moisture content of the soil varied around 20% down to 6 feet. Assuming the absence of capillary movement below the lento-capillary point, this implies that evaporation had taken place not only at the surface, but down to the level explored. The tendency for the curve connecting the moisture content of the unflooded soil to swing to the left at the top reflects the more intense evaporation at the surface.

Six days after flooding, the moisture content was fairly constant over the first two feet and then fell off gradually, the form of the curve suggesting an effect down to at least 6 feet. Subsequently the drying at the surface was rapid, but there was a lag in moisture content at lower levels so that below the first foot, even after 256 days, it was greater than before irrigation. The steadiness of the moisture content below 4 feet during the first period suggests that losses were mainly by evaporation, aided to some extent by capillary rise, from the first four feet. Thereafter, losses occurred throughout the profile.

**EFFECT OF DRAINAGE**

It would be of interest to determine, for some typical profiles, the moisture distribution attained after free drainage in the absence of losses by evaporation or gains from rainfall. In this connexion the effect of artificial drainage may also be considered.

WATER RELATIONSHIPS OF SOILS

The soil conditions leading to excessive wetness and calling for improvement by artificial drainage are of three main types:

(1) The wetness may be due to a regional water-table associated with a pervious soil profile. Such a water-table fluctuates between summer and winter level, and if a system of pipes or ditches is laid out, leading to a suitable outfall, the water-table will be lowered to a level determined approximately by the level of the outfall. The theory of drainage flow has been developed by E. C. Childs, using the analogy of two-dimensional electrical flow in a sheet conductor.

(2) The wetness may be due to the impervious character of the sub-surface horizons. The effect of this is to give a "perched" water-table which, during the wetter months of the year, may approach the surface. Where there is a natural slope to the local river system the removal of excess water will depend largely on the depth of porous soil above the impervious layer. Thin soils with impervious clay at a few inches below the surface soon become water-logged and, owing to the restriction of root range, are also liable to suffer from summer drought.

The effectiveness of a drainage system in such soils depends on the existence of pervious soil and subsoil immediately above the tiles. If drainage tiles are laid deep in impervious clay, and if the filling becomes consolidated, they cease to function. Similarly with mole drainage, the system is effective only as long as the mole is open to the soil above. Soils with this type of wetness often occur in low-lying situations and therefore in addition to the necessity for securing that the excess water shall reach the tiles or moles it is also necessary to ensure suitable gradients and outfalls.

(3) Local causes of wetness are the occurrence of springs, often where pervious upland soils adjoin lowland soils. Here the unwanted moisture can often be removed by tapping individual springs or by running a ditch or drain along the foot of the slope where the springs occur.
It is important to realize that the removal of water by artificial drainage cannot dry the soil below the moisture content corresponding with the situation of a water-table at the level of the drains. This moisture content will be rather above the field capacity, which would be attained only if the water-table were lowered to a depth considerably greater than that at which drains are usually laid.

The further drying out of wet soils can be effected only by evaporation. It is a matter of common knowledge that, under British conditions, clay soils are persistently wet in winter, even although little rain falls. The reduction of the moisture content of such soils to the point at which they can be worked is mainly dependent on the prevalence of drying winds, and cannot be effected by drainage alone. Similar considerations apply to highly organic soils.

H. H. Nicholson has shown the importance for drainage of the structure of the surface layer in heavy soils.

**LOSSES AND GAINS OF WATER UNDER FIELD CONDITIONS**

Movements of liquid water in soils are mainly due to gravitational water. The penetration of water downwards into dry soil depends on the saturation or approximate saturation of superficial layers. After the soil has been dried out to a considerable depth during summer drought, the penetration of moisture is very slow and, under British conditions, the sub-soil sometimes remains fairly dry until mid-winter.

The range of upward capillary movement is now known to be comparatively small. In Keen’s experiments, 80 cm. appeared to be the limit of upward movement. From the data of Shaw and Smith, the limit appears to be somewhat greater. Lateral capillary movement of moisture is also confined to small distances. In the absence of water percolating down slopes, soil protected from rainfall remains dry, even if surrounded by soil almost completely saturated with moisture. The soil under cattle sheds in fields is seen to be
quite dry even in mid-winter. A. H. Hendrickson and F. J. Veihmeyer found the maximum lateral movement of moisture from an irrigation trench to be under 30 inches.

Under field conditions, the absorption and disposal of rainfall depends on the texture and structure of the soil. Sandy soils rapidly absorb rainfall and allow ready percolation to the water table. Clay soils oppose a greater resistance to water movements, but differences are observable depending on structure and on the occurrence of compacted horizons. Ferrallitic clays are more open in structure and thus less subject to surface washing under heavy rainfall than aluminosilicic clays, which may not be able to dispose of intense rainfall sufficiently rapidly to prevent run-off and consequent erosion. Other things being equal, the most siliceous clays show the greatest tendency to dispersion, resulting in drainage impedance.

Owing to continued interference through gains from rainfall on the one hand and losses by evaporation on the other, the attainment of moisture equilibrium in a soil profile must be regarded as a purely ideal condition. It is of interest, however, to consider the type of moisture distribution that might be expected in a soil profile with a water-table at such a depth as to be without influence on surface conditions.

During a period when rainfall predominates over evaporation, the tendency will be towards a distribution in which the moisture content at each level will approximate to that represented by the field capacity. This does not necessarily imply an uniform distribution, since the field capacity will vary with the colloidal content and also, to some extent, with the structure. Uniform distribution would be approached only in a uniform soil. Excess of moisture over this ideal distribution, owing to recent precipitation, would be reflected in a higher moisture content at the surface. In the absence of evaporation, this would be translated to lower levels partly by gravitational movement and partly by capillary transmission and would increase the depth of soil
moistened to the field capacity. If the depth thus moistened extends to the zone affected by the water-table, any excess of moisture would be handed on to the ground-water. The rapidity with which such an adjustment is attained will depend on the permeability of the soil to water movements. In heavy soils, adjustments are slow and moisture contents considerably above the equilibrium distribution may persist in the surface horizons for long periods.

With soils of low permeability, heavy rain may result in the saturation of the surface soil. Further precipitation is then disposed of by superficial run-off, with the possibility of erosion of the top soil. The susceptibility of soils to erosion in this way depends on their texture, structure, organic matter content, and the composition of the clay complex. Generally speaking, the tendency to a deflocculated or single-grain structure implies liability to erosion by run-off.

We may now consider the changes in moisture distribution consequent on the cessation of precipitation and the onset of conditions favouring evaporation. If the surface soil is saturated above its equilibrium moisture content by recent rain, losses will take place both by evaporation and also by capillary and gravitational translation to lower levels. Losses by movement to lower levels will cease when the equilibrium moisture distribution has been reached, and thereafter, losses will be entirely by evaporation, at first mainly from the surface, but subsequently, as in Greene's experiments, from the lower horizons. In the earlier stages of drying-out, there is also the possibility of upward capillary movements, but as each successive layer becomes dried to the lento-capillary point, the range of upward movement will be restricted and the losses by evaporation from the superficial horizons will no longer be replaced from below. Since the rate of direct evaporation is greatest at the surface and decreases with depth, the rate of loss of moisture from the soil as a whole must, ceteris paribus, decrease when capillary movement to the surface is no longer possible, and will be determined mainly by the direct evaporation from the surface and the subjacent layers.
Where a water-table is present near the surface, conditions will be different, for capillary rise will suffice to replace the losses by evaporation from the surface. But, as we have seen, when the water-table is below a depth of a few feet capillary rise can no longer cover the losses by evaporation from the surface layers. The soil then dries out from the surface downwards.

The drying-out process is much more rapid under a cover of vegetation than with bare soil. E. S. West, in New South Wales, found that with irrigated crops (lucerne, cowpeas, Bokhara clover) the water-table was lowered considerably below that under bare soil, although the crops received 3 to 4 times as much water by irrigation and rainfall. Lucerne is particularly effective in drying out the soil. It is possible that a change in the character of the vegetation, for example, by afforestation or the growth of forage crops, might be effective in aiding the removal of excessive moisture from soils in wet districts.

It has been suggested by A. F. Lebedev that, apart from movements of water in the fluid state, a certain amount of vertical translocation takes place in the vapour state. During the winter months, the surface soil is colder than the subsoil, and this may result in an upward distillation conformable with the difference in vapour pressure. Accession of moisture in this way may be a contributory cause of the persistent wetness of soils in late winter when the temperature gradient to the surface is steepest. On the other hand, during the summer months, a downward distillation may take place from the warm surface soil to the colder subsoil, thus reinforcing the losses by evaporation from the surface.

BIBLIOGRAPHY


CHAPTER X

SOILS OF THE PODSOLIC GROUP

INTRODUCTORY

In this and the ensuing chapters, examples will be given of the soils of the principal world-groups. In arranging the material into chapters, the writer is aware that he is adopting a classification before the general principles underlying soil classification in general have been discussed. The alternative arrangement of discussing the principles of classification and then proceeding to the description of soil groups was considered; but it was felt that such a discussion could be more easily undertaken after the reader had made some acquaintance with the material to be classified. In any case, the titles used as headings for this and the following chapters correspond, in most systems of classification, with the soils assembled under them.

The soils to be described in this chapter occur under humid conditions and are best developed in cold to temperate regions. The podsols* and their congeners are completely leached soils in which calcium carbonate and calcium sulphate are present only as fugitive constituents. The reaction is thus on the acid side of neutrality. The natural vegetation is generally forest or heath, but prairie vegetation is associated with one of the types. The soils of swamp, fen, and marsh vegetation, which appear as local associations

*The alternative spelling podsol is generally used in the United States and, I understand, corresponds with the original Russian. After some consideration, I have decided to retain the spelling that has now become familiar to most English readers and is sanctioned by usage.
where drainage conditions are sufficiently impeded, belong to the hydromorphic soils to be discussed in Chapter XII.

PODSOLS

This group of soils, by reason of its wide extent in Russia and Northern Europe, has received more detailed attention than any other group of soils, with the possible exception of the chernozems.

The podsols of the northern hemisphere occupy the humid regions lying to the south of the tundra (p. 352). With their related types, they occupy the greater part of northern and western Europe. They are developed typically under coniferous forest, but occur also under heath vegetation which, in its meagre demands on the plant nutrients of the soil, exhibits a close resemblance to coniferous forest.

A typical podsol profile has already been described in Chapter IV. (p. 84) and consists essentially of three horizons. The A, or eluviated horizon, consists of a layer of peaty material underlain by a more or less bleached layer, relatively poor in humus and sesquioxides. Below the A horizon is the B horizon. This is enriched by certain of the constituents leached from the A horizon. The constituents that distinguish this horizon by their accumulation are humus and sesquioxides. As the conditions of deposition vary with the relative proportion of humus and sesquioxides in the percolating moisture and also with its reaction, different kinds of podsol may be distinguished. In the more extreme types, the humus podsols, the B horizon is enriched by an accumulation of humus and sesquioxides, the accumulation of sesquioxides occurring at a lower level in the profile than that of the humus. In less extreme podsols, there may be no accumulation of humus, but an accumulation of hydrated ferric oxide in the B horizon. These are known as iron podsols. The C horizon is the parent material from which the profile is developed. The precipitation of
materials in the B horizon of podsol soils often leads to the formation of hardpan (ortstein). Such a condition may be followed by impediment of drainage and a change in the character of the soil profile to gley-podsol (p. 356).

The dominant factor in the development of the podsol profile is the prevalence of intense leaching owing to the continued excess of rainfall over evaporation.

Whilst podsol soils may be developed on all kinds of parent materials, the degree of development is affected by the base-status of the substrate and by its permeability to leaching. The most strongly developed profiles occur in quartzose sands, which are naturally poor in basic constituents. In such cases, even where the climate is not of the most extremely humid type, impoverishment rapidly occurs and the only possible type of vegetation consists of plants such as conifers and heath plants that can exist on minimal supplies of mineral plant food. With poverty in basic constituents, the residues of vegetation undergo a type of decomposition, chiefly through the action of fungi, resulting in the production of a layer of raw humus which, in the almost complete absence of earth-worms, remains sharply differentiated from the mineral soil over which it lies. The extremely acid leaching results in the development of a deep bleached layer and a humic B horizon. Such podsol soils may be seen on the Bagshot Sands of S.E. England. They are widespread on the sandy soils of Northern and Middle Europe.

Podsol soils are less readily developed in materials with high base reserves than in light quartzose sands. In loams and clays, the podsolized layer is shallower and indeed, in some cases, cannot be recognized without chemical analysis of the colloidal material. It is obvious, then, that the distribution of podsol soils is markedly influenced by geology. If the whole of England were a region of light sands, such as the Bagshot Beds, it would be represented on a soil map as an area of podsol soils. Actually, owing to the prevalence of clays and loams having notable base reserves, well developed
podsols are by no means common. It is possible that in certain upland regions, original podsols have been truncated by erosion consequent on deforestation in former centuries. In such areas, secondary podsols may be developed in the original B horizon, which has been laid bare by erosion.

It is unfortunate that the available data for podsol profiles cannot be uniformly presented. Many continental and American workers have described profiles in terms of the total analyses of the individual horizons. Such figures, in the writer's opinion, are not highly instructive since they include both the unweathered skeleton of the soil and the weathering-complex, which is the seat of the chemical reactions of the soil. And thus an enrichment in sesquioxides in a given horizon, as shown by the total analyses, may either reflect the mechanical washing down of colloidal clay, or the translocation of sesquioxides from the clay complex of an eluvial horizon and their deposition in the horizon in question. The most desirable information would be in the form of chemical and mineralogical data for the composition of the clay complex in the different horizons of a profile. This information is given to some extent by certain workers who report constituents soluble in hydrochloric acid. If it could be assumed that the acid extraction used involved no attack on the unweathered minerals of the soil skeleton, such data would present a comparative view of the composition of the clay complex in the different horizons and would furnish information as to the character of the eluviation processes. Few soils, however, are free from original minerals susceptible to attack by hydrochloric acid. The error from this complication is likely to be greatest in relatively young soils. With certain tropical soils and soils whose parent materials have passed through long cycles of weathering, such as many of the Tertiary and Recent sediments, the proportion of unweathered silicates is likely to be less and the soil skeleton will consist mainly of quartz. Here the figures obtained by the use of hydrochloric acid extractions may have a real significance.
Russia.* A sandy profile in the region of St. Petersburg (Leningrad) was described by Georgievski¹ as follows:—

Horizon A₁. Loose greyish-white layer, 4-5 inches thick.
Horizon A₂. White fine sand 10 inches or more thick.
Horizon B. A more or less compact or even hard mass of brown or black colour. In parts less cemented; thickness 5-10 inches.
Horizon C. Reddish-yellow or yellowish glacial sand.

In the same district, a clay soil gave:—

Horizon A₁. Whitish-grey fine grained layer, 5-6 inches thick.
Horizon A₂. Compact laminated almost white material, appearing white when dry and falling into a fine mealy powder. Some loam concretions present; thickness 3-5 inches.
Horizon B. Compact clayey material with abundant dark and brown concretions. The colour is mottled, whitish flecks alternating with reddish and yellowish streaks and threads of the slightly altered parent material; thickness about 8 inches.
Horizon C. Compact boulder clay of reddish-brown colour.

It will be noticed that in a heavier parent material the profile is markedly shallower. The tendency to laminated structure in the A₂ horizon is also to be noted and is a common characteristic of well developed podsols.

It is to be presumed that, in the profiles just described, the B horizon is enriched both in organic matter and in sesquioxides. In certain sandy and gravelly profiles, the humus occurs as a distinct layer above the horizon of deposition of the sesquioxides.

Denmark. Podsol profiles under heath formation in Jutland, Denmark, have been exhaustively investigated by

*The profiles described in this and the ensuing chapters have been chosen to illustrate the distinctive features of the groups to which they belong. No attempt has been made to give examples from all countries. In the descriptions given, the horizons A, A₁, etc., are usually given as in the papers from which they are quoted. In modern profile descriptions it is customary to use $A_{00}$ to denote litter, $A₀$ raw humus, $A₁$ mineral soil with humus infiltration, $A₂$ bleached horizon, $B₁$ horizon of illuvial humus deposition, $B₂$, $B₃$, etc., horizons enriched by deposition of sesquioxides. A discussion of the morphology of the horizons of the podsol profile is given by J. S. Joffe (Soil Sci., 1932, 33, pp. 217-238).
I. Profile from Goldstone, Salop, England.

II. Profile in Glacial Till, Minnesota, U.S.A.

PLATE IV.
F. Weis². Varying degrees of podsolization are reflected in the changes in the composition of the inorganic colloids extracted by the ammonium oxalate method of O. Tamm³. Marked accumulation of humus with hardpan formation is shown in the B horizons of the most strongly podsolized profiles. Sesquioxide accumulation is most marked in the humus horizon, as shown by the composition of the ammonium oxalate extract.

Sweden. K. Lundblad⁴ has described and examined a number of iron podsols in Sweden. The following profile occurs at Siljanfors under a sparse pine wood with Calluna vulgaris, Vaccinium vitis-idaea, V. myrtillus, and lichens as ground vegetation:—

A₀. 1 cm. Duff, a loose organic (F) layer.
A₁. 7 cm. Pinkish-white bleached layer.
B. 27 cm. Reddish-yellow, darkest at top.
C. Morainic material mainly sandstone material with some granite.

Laboratory examination by Tamm’s acid oxalate method shows an enrichment in sesquioxides and also in silicic acid in the B horizon.

S. Mattson and Y. Gustafsson⁵ have also examined Swedish podsols from the standpoint of the former’s theories of isoelectric precipitation. They find the ultimate pH to be highest in C, lower in B, and lowest in A horizons. In the B and C horizons the actual pH is below the ultimate pH.

England. W. M. Davies and G. Owen⁶ give a description of a podsol profile at Goldstone, Newport, Shropshire, developed on Bunter pebble beds under a vegetation of scrub oak, birch, bilberry, ling, heather, etc. This profile is illustrated in Plate IV. Their description is as follows:—

A₀. 0-6 in. Leaf mould.
A₁. 6-9 in. Greyish-black humified organic matter with bleached sand grains and numerous pebbles.
A₂. 9-19 in. Pale grey loose sand with bleached sand grains. Numerous pebbles free of iron oxide stains.
B₁. 19-23 in. Humified organic matter compacted.
B₂. 23-35 in. Compact cinnamon-yellow light sand with numerous stained pebbles. Decreasing intensity of coloration with depth. Upper surface consists of 1" layer of cemented and lustrous iron oxide.

C. Below 35 in. Pebbly brownish laminated and friable sandy rock.

The principal analytical data are shown in Table XXVII.

**Table XXVII.—Analytical Data for Goldstone Podsol Profile (Davies and Owen)**

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Clay %</th>
<th>Organic Carbon %</th>
<th>pH</th>
<th>Composition of clay fraction</th>
<th>SiO₂/Al₂O₃</th>
<th>SiO₂/R₂O₃</th>
<th>Al₂O₃/Fe₂O₃</th>
</tr>
</thead>
<tbody>
<tr>
<td>A₁(6-9&quot;)</td>
<td>2.7</td>
<td>14.88</td>
<td>3.70</td>
<td>3.13</td>
<td>2.55</td>
<td>4.43</td>
<td></td>
</tr>
<tr>
<td>A₂</td>
<td>1.3</td>
<td>1.63</td>
<td>3.13</td>
<td>2.81</td>
<td>2.66</td>
<td>18.00</td>
<td></td>
</tr>
<tr>
<td>B₁</td>
<td>8.8</td>
<td>7.14</td>
<td>3.73</td>
<td>1.79</td>
<td>1.66</td>
<td>13.03</td>
<td></td>
</tr>
<tr>
<td>B₂</td>
<td>2.2</td>
<td>0.40</td>
<td>3.75</td>
<td>1.51</td>
<td>1.19</td>
<td>3.80</td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>3.3</td>
<td>0.04</td>
<td>4.75</td>
<td>2.29</td>
<td>1.94</td>
<td>5.54</td>
<td></td>
</tr>
</tbody>
</table>

The above data give a clear picture of the process of podsolization. In the first place, there is some evidence of mechanical eluviation in the rise in clay content in the B₁ horizon. Secondly, there is a marked rise in the organic carbon content in the B₁ horizon. Thirdly, the siliceous character of the clay fraction in the A horizons and the contrasted sesquioxidic character of the clay fraction in the B horizons should be remarked. Finally, the strongly acid reaction of the whole profile, and the increase in pH with depth is shown.

**Scotland.** Podsol profiles are of frequent occurrence in the Scottish Highlands, where they are developed under heath or coniferous forest on igneous or metamorphic parent materials. A typical example on granitic drift in Aberdeenshire is described by A. B. Stewart⁷.

The mean annual rainfall is 34 inches and the vegetation *Calluna vulgaris* dominant, with *Vaccinium myrtillus* and
Erica cinerea, and conifers replanted after de-forestation. The description is as follows:—

1. 0-2 cm. Surface litter.
2. 2-8 cm. Dark brown to blackish layer with white specks of SiO₂; slightly laminar in structure; plant-remains decipherable and roots abundant.
3. 8-18 cm. Sandy loam; light grey mineral particles and brown organic material; structureless; plastic but gritty; plant-remains decipherable and roots present.
4. 18-30 cm. Light brown sandy loam; friable and permeable; organic matter well decomposed; roots present; contains scattered vertical intrusions of surface material.
5. 30-55 cm. Dark brown, coarse, sandy loam; friable but more compact than (4); organic matter well decomposed; roots stop sharply at bottom of this layer.
6. 55 cm. + Yellowish-fawn sand and gravel with horizontal rusty streaks; cemented but quite friable when broken; stones increase in quantity and size with depth.

The principal analytical data are given in Table XXVIII.

**TABLE XXVIII.—ANALYTICAL DATA FOR SCOTTISH PODSOL (STEWART)**

<table>
<thead>
<tr>
<th>Depth (in cm.)</th>
<th>2-8</th>
<th>8-18</th>
<th>18-30</th>
<th>30-55</th>
<th>55+</th>
</tr>
</thead>
<tbody>
<tr>
<td>pH</td>
<td>5.2</td>
<td>4.5</td>
<td>4.9</td>
<td>5.0</td>
<td>5.8</td>
</tr>
<tr>
<td>Ignition Loss</td>
<td>44.4</td>
<td>12.0</td>
<td>13.8</td>
<td>15.8</td>
<td>4.6</td>
</tr>
<tr>
<td>Clay %</td>
<td>5.7</td>
<td>10.4</td>
<td>5.4</td>
<td>1.5</td>
<td></td>
</tr>
<tr>
<td>Clay SiO₂/Al₂O₃</td>
<td>2.84</td>
<td>2.07</td>
<td>1.28</td>
<td>1.29</td>
<td></td>
</tr>
<tr>
<td>&quot; SiO₂/R₂O₃</td>
<td>2.46</td>
<td>1.44</td>
<td>0.91</td>
<td>1.04</td>
<td></td>
</tr>
<tr>
<td>&quot; Al₂O₃/Fe₂O₃</td>
<td>6.47</td>
<td>2.36</td>
<td>2.46</td>
<td>4.16</td>
<td></td>
</tr>
</tbody>
</table>

The data show some eluviation of clay and sesquioxides, but do not suggest any humus B development. The horizons to 18 cm. are of an A character, those from 18-55 cm. are B horizon, whilst below 55 cm. is C material.

Wales. The following profiles are typical.

Cefnybryn, W. Glamorgan. Altitude 300 feet. Mean
annual rainfall, 40 inches. Mean annual temperature, 50°F. Parent material, Old Red Sandstone Conglomerate. Vegetation, *Ulex spp.*, *Festuca spp.* (described by D. O. Hughes and Brynmor Jones*).

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Depth</th>
<th>Organic Carbon %</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>A.</td>
<td>0-4 in. Fibrous peat layer</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>A.</td>
<td>4-7 in. Greyish sand</td>
<td>...</td>
<td>1.61</td>
</tr>
<tr>
<td>B.</td>
<td>7-12 in. Reddish-brown sandy loam</td>
<td>...</td>
<td>2.08</td>
</tr>
<tr>
<td>C.</td>
<td>14-16 in. Yellowish-brown sandy loam on rock</td>
<td>...</td>
<td>0.48</td>
</tr>
</tbody>
</table>

The composition of the clay fraction in the several horizons is as follows:

<table>
<thead>
<tr>
<th>Horizon</th>
<th>SiO₂/Al₂O₃</th>
<th>SiO₂/R₂O₃</th>
<th>Al₂O₃/Fe₂O₃</th>
</tr>
</thead>
<tbody>
<tr>
<td>A.</td>
<td>3.70</td>
<td>3.33</td>
<td>9.09</td>
</tr>
<tr>
<td>B.</td>
<td>2.44</td>
<td>1.67</td>
<td>2.17</td>
</tr>
<tr>
<td>C.</td>
<td>2.80</td>
<td>2.03</td>
<td>2.63</td>
</tr>
</tbody>
</table>

The data indicate a slight accumulation of organic matter and alumina and a marked accumulation of ferric oxide in the B horizon.

Aber, Caernarvonshire. Altitude, 900 feet. Mean annual rainfall, about 60 inches. Mean annual temperature, 47°F. Parent material, Ordovician shale. Vegetation, *Calluna vulgaris, Vaccinium myrtillus, Festuca spp.*, *Lichens.* (Described by G. H. Gethin Jones*.)

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Depth</th>
<th>Organic Carbon %</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>A.</td>
<td>0-5 in. Dark peaty layer</td>
<td>...</td>
<td>13.8</td>
</tr>
<tr>
<td>A.</td>
<td>5-8 in. Grey stony loam</td>
<td>...</td>
<td>6.2</td>
</tr>
<tr>
<td>B.</td>
<td>8-14 in. Reddish-brown loam</td>
<td>...</td>
<td>4.7</td>
</tr>
<tr>
<td>B.</td>
<td>14-20 in. Light brown shaly loam</td>
<td>...</td>
<td>3.5</td>
</tr>
<tr>
<td>B.</td>
<td>20-26 in. Light brown very shaly loam</td>
<td>...</td>
<td>3.6</td>
</tr>
</tbody>
</table>

The composition for the clay fraction in the several horizons is as follows:

<table>
<thead>
<tr>
<th>Horizon</th>
<th>SiO₂/Al₂O₃</th>
<th>SiO₂/R₂O₃</th>
<th>Al₂O₃/Fe₂O₃</th>
</tr>
</thead>
<tbody>
<tr>
<td>A.</td>
<td>4.40</td>
<td>3.56</td>
<td>4.24</td>
</tr>
<tr>
<td>A.</td>
<td>3.84</td>
<td>2.75</td>
<td>2.52</td>
</tr>
<tr>
<td>B.</td>
<td>1.90</td>
<td>1.31</td>
<td>2.22</td>
</tr>
<tr>
<td>B.</td>
<td>2.09</td>
<td>1.39</td>
<td>1.99</td>
</tr>
<tr>
<td>B.</td>
<td>1.56</td>
<td>1.17</td>
<td>3.00</td>
</tr>
</tbody>
</table>
The data give no indication of accumulation of organic matter, but clearly show the deposition of ferric oxide and, to a smaller degree, alumina in the B horizons.

**United States.** Podsols are of frequent occurrence throughout the more humid regions of the Northern United States and Canada. A profile collected by W. J. Geib\(^{10}\) near Bayfield, Wisconsin, is described as follows:—

A. 0-3 in. Dark brown decayed leaves and other organic matter.
A. 3-8 in. Grey with very slight pink tint.
B. 12-30 in. Light to dark coffee brown, compactly cemented.
C. 30-40 in. Pinkish-grey resting on heavy red clay.

The profile is developed in a fine sandy loam and shows a definite accumulation of clay in the B horizon. The silica-sesquioxide ratios of the colloidal clay in the successive horizons are 3.76, 4.01, 1.59, and 2.40 respectively.

W. J. Latimer\(^{11}\) describes the following profile from Washington, Massachusetts:—

A. 0-6 in. Dark brown with mould and roots.
A. 6-11 in. Essentially grey, but containing brown spots.
B. 11-13 in. Dark brown or coffee coloured.
C. 24-36 in. Yellowish-grey and streaked mottled brown.

The texture is a loam and increases in heaviness in the B and C horizons. The B\(_1\) horizon shows an accumulation of organic matter. The silica-sesquioxide ratios of the colloidal clay in the successive horizons are 2.16, 2.74, 0.86, 1.28, and 1.67, respectively.

M. M. McCool, J. O. Veatch, and C. H. Spurway\(^{12}\), in studies on soil profiles in Michigan, have given the following generalized description of podsolized soils from the northern part of the State.

1. Thin humous soil (mould).
2. Grey podsolized soil (3-24 in.).
3. Brown horizon (dark coffee brown to light leather colour and dull yellow); thickness 4 in. to 4 feet; horizon of acid concentration.
4. Horizon showing iron oxide coloration; highest clay content; gradation to substratum.
5. Substratum.
H. A. Lunt gives a description of New England soils under different types of forest vegetation and humus cover. They are generally of the iron podsol type and the degree of development is closely correlated with the type of humus cover. A Minnesota podsol is shown in Plate IV.

Tropics. It was formerly considered that podsol soils were confined to temperate and cold humid regions. According to P. Vageler, however, podsolization can occur to some extent in the tropics under conditions of extreme humidity. H. J. Hardon has described certain soils in S.E. Borneo, known as padang soils, which are definitely podsol. They occur at about 90 metres altitude under 3,000 mm. annual rainfall, and mean annual temperature 26°C. The vegetation is of a heathy type and the parent material is a very coarse sand. The following is a typical profile.

\[ \begin{align*}
A_0 & \quad 0-20 \text{ cm.} \quad \text{Brown-black layer of semi-decomposed plant residues intermixed with coarse quartz sand.} \\
A_1 & \quad 10-25 \text{ cm.} \quad \text{Loose greyish-black humic quartz sandy layer.} \\
A_2 & \quad 25-40 \text{ cm.} \quad \text{Loose greyish-white quartz sandy layer.} \\
B_1 & \quad 40-70 \text{ cm.} \quad \text{Dark brown very compact quartz sandy hardpan.} \\
B_2 & \quad 70-100 \text{ cm.} \quad \text{Loose light brown quartz sandy layer.}
\end{align*} \]

The silica-sesquioxide ratios of the clay in the horizons are as follows:

\[
\begin{align*}
A_0 & \quad 3.62 \\
A_1 & \quad 7.17 \\
A_2 & \quad 8.64 \\
B_1 & \quad 0.42 \\
B_2 & \quad 0.31
\end{align*}
\]

TRUNCATED PODSOLS

In hilly districts, soils are frequently encountered in which the A horizon has apparently been removed by erosion. G. W. Robinson describes such a profile at Aber, Caernarvonshire. It occurs under grass and bracken under a mean annual rainfall of 60 in. at 800 feet.

0-3 in. Turfy layer.
3-6 in. Dark brown turfy loam.
6-9 in. Brown loam.
9-12 in. Light brown, light loam.
12-15 in. reddish-brown shaly loam on Ordovician shale.
The principal analytical data are shown in Table XXIX.

**Table XXIX. — Analytical data for truncated podsol**

<table>
<thead>
<tr>
<th>Depth</th>
<th>0-3 in.</th>
<th>3-6 in.</th>
<th>6-9 in.</th>
<th>9-12 in.</th>
<th>12-15 in.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Organic carbon %</td>
<td>...</td>
<td>15.9</td>
<td>8.1</td>
<td>5.4</td>
<td>3.4</td>
</tr>
<tr>
<td>pH</td>
<td>...</td>
<td>4.20</td>
<td>4.42</td>
<td>4.58</td>
<td>4.78</td>
</tr>
<tr>
<td>Clay fraction: SiO₂/Al₂O₃</td>
<td>2.17</td>
<td>2.25</td>
<td>2.17</td>
<td>1.95</td>
<td>2.12</td>
</tr>
<tr>
<td>&quot;</td>
<td>SiO₂/Fe₂O₃</td>
<td>1.63</td>
<td>1.69</td>
<td>1.62</td>
<td>1.48</td>
</tr>
<tr>
<td>&quot;</td>
<td>Al₂O₃/Fe₂O₃</td>
<td>3.02</td>
<td>3.02</td>
<td>2.95</td>
<td>3.13</td>
</tr>
</tbody>
</table>

The B character of the surface soil may be even more pronounced in other examples. SiO₂/Fe₂O₃ ratios of less than 1.0 have been recorded at higher altitudes.

**VARIATIONS IN PHYSICAL PROPERTIES IN PODSOL PROFILES**

The vertical variation in the physical properties of podsol profiles is governed by the texture of the parent material, the degree of development of the profile, and the thickness of the peaty layer. A. F. Lebedev and E. E. Bankova have followed the vertical changes in a number of physical constants, including volume weight, porosity, and maximum molecular water-holding capacity for representative profiles. In general, the volume weight increases with the depth, but whilst in some cases it reaches a maximum in the B horizon, in other cases there is a continuous increase from the surface to the C horizon. The variations in porosity are generally in the opposite direction, a minimum occurring sometimes in the B, sometimes in the C horizon. Maximum molecular water capacity may increase with depth, and pass through a maximum in the B horizon or, in the case of humus podsol, show a maximum in the upper A horizon.

There is generally a certain amount of mechanical eluviation in podsol profiles. The accumulation of colloidal material in the B horizon is most marked in soils of light
texture. This accumulation may be so pronounced in some cases as to lead to a certain amount of impedance of drainage with consequent waterlogging—a condition encountered in some of the podsol soils of the Bagshot area of Surrey. The development of an horizon with restricted permeability is doubtless a factor in the precipitation of sesquioxides and the development of a pan. At a certain stage of development of pan the restriction of drainage may be sufficient to change the character of the profile and gley podsols may be formed, passing ultimately into peat podsols and peats.

The degree of induration of the B horizon varies considerably. In some cases, although there is a definite horizon of accumulation, induration is only slightly developed. At the other extreme are rock-like B horizons such as occur in the Bagshot profiles. A pebbly or stony texture appears to favour the formation of highly indurated illuvial horizons.

In podsols developed from substrates of heavier texture, such as the shaly loam profiles of Wales, mechanical eluviation is less marked and the induration in the B horizon is comparatively slight.

MECHANISM OF PODSOLIZATION

We have already seen that podsolization is a consequence of the removal of bases by leaching and the development of an acid organic layer at the surface. Sulphuric acid formed by the biological oxidation of organic sulphur compounds may also play a part in the leaching processes (cf. p. 218). With regard to the mechanism of the process, some difference of opinion exists. The accumulation of sesquioxides in the B horizon may reflect either a translocation and deposition of clay, which, relative to the coarser fraction of the soil, is rich in these constituents, or an actual differentiation of the clay complex due to a simple translocation of sesquioxides from the A to the B horizons. Whilst it is certain that in some cases mechanical eluviation, resulting in an actual transference of inorganic colloidal material from
the A to the B horizon, takes place, the differences found in
the composition of the clay complex in the several horizons
by certain workers can only be explained by postulating a
chemical eluviation.

The translocation and deposition of humus occurs only
under very acid conditions. B. Aarnio has studied the
mutual precipitation of humus and sesquioxide soils and finds
that with ferric oxide-humus ratios varying from 1:0.2 to
1:2.8, precipitation occurs. The corresponding range for
mutual precipitation of humus and alumina soils is 1:1 to
1:30. It can thus be seen that ferric oxide is more mobile
in the soil profile than alumina, for when humus is in excess
in the soil solution, at least 30 times as much humus as
alumina must be present for this constituent to pass into
colloidal solution. A more moderate concentration of
humus suffices to ensure the translocation of colloidal ferric
oxide. It is to be presumed that soils having the sesqui-
oxides in excess of the humus do not occur under ordinary
conditions. Assuming a sufficient solubility of the humus to
give a humus-alumina ratio of over 30, then a decrease in
this ratio owing to the removal of some of the humus will
result in a mutual precipitation of the humus and alumina.
Sufficient humus will remain in solution to protect the ferric
oxide, which will be precipitated at lower levels.

The differentiation of the clay complex implied by the
variations in its composition in the horizons of a normal
podsol profile is generally held to be a consequence of its
instability under conditions of base-unsaturation. Where
there is base-saturation, the clay complex remains stable and
no differentiation occurs. An essential for podsolization
is the presence of raw humic matter such as occurs under
coniferous forest or heath. It would appear that an acid
reaction is not in itself sufficient to cause podsolization.
This may be readily shown in the laboratory by allow-
ing 0.1 N-solutions of hydrochloric acid and oxalic acid,
respectively, to percolate through ferruginous sand.
Although the pH of the hydrochloric acid is lower than
that of the oxalic acid, the solvent action of the latter acid is considerably greater and quickly leads to a bleaching by removal of ferric oxide. H. T. Jones and J. S. Wilcox\textsuperscript{19}, from laboratory studies of the solution of sesquioxide in A horizons and their precipitation in B horizons, conclude that hydroxy-acids play a part in the solution of sesquioxides, which are translocated in combination as complex anions and precipitated as basic salts. P. H. Gallagher\textsuperscript{20} considers that the effective agent in podsolization is not the hydrogen ion but the associated anions principally of hydroxy-acids and oxalic acid formed by the decomposition of plant residues. According to this view acidity is not necessary for podsolization, although it is almost invariably present. Certain Irish podsolic soils have been found to occur with neutral reaction and high base-status.

The deposition of sesquioxides removed by leaching from the A horizon in podsol profiles has not yet been completely explained. There is generally an increase in base-saturation with depth, and this in itself would lead to some deposition of constituents which have been brought into solution by acid percolating waters. There is also the possibility that accumulation of sesquioxides may result, in part, from the restriction of free percolation in an horizon enriched in clay by mechanical eluviation.

C. G. T. Morison and D. B. Sothers\textsuperscript{21} consider the deposition of ferric oxide in iron pans to be due to irreversible precipitation of iron from humus-protected iron sols during periods of drought. During the winter months, the movements of soil moisture are mainly downwards and this results, if there is a surface layer of acid humus, in a downward leaching of ferric oxide, and, possibly, of alumina, in colloidal solution. With the drying out of the profile during summer, precipitation occurs. The localization of this precipitation in a definite horizon is possibly connected with the course of the moisture gradient. It is known that capillary movement becomes negligible at moisture contents in the vicinity of the moisture equivalent. Morison and
SOILS OF THE PODSOLIC GROUP

Sothers consider the horizon of precipitation to be determined by the proximity of a level of complete saturation, but it does not appear possible to postulate a water-table so near the surface. The horizon of deposition may occur at a depth where the subjacent strata are sufficiently moist to furnish a supply of precipitable material by capillary rise.

This view of the deposition of sesquioxides may apply also to the deposition of humic material. If it is correct, the B horizons are developed by precipitation not from descending solutions during periods of leaching but from ascending solutions during periods of drought.

The rate of podsolization varies enormously with climate, parent material and vegetation. It is not, however, a process requiring "geological time." O. Tamm estimates that in N. Sweden, under Scots pine, 1-2 cm. of bleached layer is formed in a century.

An exhaustive discussion of the process of podsolization is given by J. S. Joffe. S. Mattson and Y. Gustafsson have discussed podsolization from the standpoint of the former's iso-electric complex theory.

CRYPTOPODSOLIC SOILS

The term cryptopodsols, as distinct from phaneropodsols, is used by C. C. Nikiforoff to describe soils in which the podsolic character is demonstrable only by chemical analysis. The soils represent a transition to the soils described in the next section in which sesquioxide translocation is recessive. They occur abundantly in Wales and are characterized by a dark humus A horizon overlying a B horizon which is generally yellowish or orange-brown in colour. Examination of the clay fraction reveals a definite increase in sesquioxides in passing from A to B. On oxidizing with hydrogen peroxide to destroy humus it is generally observed that the A horizon is somewhat greyer in colour than the B horizon. The slight podsolization is therefore masked by the humus colour.
The following profile may be cited as an example of soils of this class:


I. 0-8". Dark greyish-brown rather stony light loam.
II. 8-12". Bright yellowish-brown rather stony light loam.
IV. 18"+. As above but more stony, passing to rock brash.

The following are the principal analytical figures:

<table>
<thead>
<tr>
<th></th>
<th>Clay %</th>
<th>Organic Carbon</th>
<th>pH</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>SiO₂/K₂O</th>
<th>Al₂O₃/Fe₂O₃</th>
<th>SiO₂/Al₂O₃</th>
</tr>
</thead>
<tbody>
<tr>
<td>I.</td>
<td>18.3</td>
<td>3.55</td>
<td>5.51</td>
<td>2.57</td>
<td>1.93</td>
<td>3.01</td>
<td></td>
<td></td>
</tr>
<tr>
<td>II.</td>
<td>18.6</td>
<td>1.43</td>
<td>5.28</td>
<td>2.16</td>
<td>1.62</td>
<td>3.00</td>
<td></td>
<td></td>
</tr>
<tr>
<td>III.</td>
<td>15.0</td>
<td>0.53</td>
<td>5.40</td>
<td>2.17</td>
<td>1.72</td>
<td>3.82</td>
<td></td>
<td></td>
</tr>
<tr>
<td>IV.</td>
<td>14.0</td>
<td>0.22</td>
<td>5.51</td>
<td>2.67</td>
<td>2.09</td>
<td>3.60</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The figures for the composition of the clay fractions reveal a definite accumulation of sesquioxide in layers II. and III.

The Spanish soils described by E. H. Del Villar as acid-humic soils would appear to fall into the cryptopodsol class for the bleaching in the A horizon, if present, is masked by humus accumulation.

BROWN EARTHS OR GREY-BROWN PODSOLIC SOILS

The name brown earths (Braunerde) was applied by Ramann to a group of soils occurring in Western Europe and developed naturally under deciduous forest. They are also known as brown forest soils and appear to correspond with the grey-brown podsolic soils of the United States. Soils of a character generally similar to the brown earths of Europe occur in many parts of the world and the attempts to correlate them have given rise to considerable discussion on the definition of the group. The following appear to be the essential characters of the great world group of which the brown earths or brown forest soils found in different parts of the world are variants.
(1). The profile is completely leached of carbonates. Carbonates may be present in the C horizon, or in the top soil as residues from added dressings. Drainage is free.

(2). The colloidal complex is not highly base-un- saturated and the reaction is only moderately acid.

(3). The humus is of the "mild" type, well distributed throughout the upper horizons without any tendency towards the development of a raw humus layer.

(4). The composition of the clay complex in soils of primary weathering tends towards a $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio of 2.0, representing a stage in desilicification intermediate between that of the chernozem group and that of the ferrallitic soils of the tropics.

(5). Free sesquioxides are present, and the hydrated ferric oxide thus occurring gives a brownish or reddish-brown colour, which may be masked by humus.

(6). There is no differential eluviation of silica or sesquioxides and the $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio of the clay remains fairly constant down the profile.

(7). The soil structure is moderately granular.

(8). The natural vegetation is deciduous woodland or scrub.

The affinity of these soils with the podsols consists in the complete leaching of carbonates from the mature soil profile and the development of a brown colour suggestive of the B horizon of the podsols. They differ from the podsols in some important respects, but chiefly in their better base-status. This may be due to either or both of two causes. On the one hand, the climate may be such as to involve a less intense leaching, whilst, on the other hand, the parent material of the soil may be sufficiently rich in basic reserves to oppose the extreme impoverishment which favours podsol development.

The return to the soil of leaf-fall containing basic constituents is a powerful factor in maintaining the base-status of these soils in their virgin state. The presence of a ground-vegetation also assists in promoting the circulation of mineral
nutrients through a deep layer of soil and subsoil. The soil under deciduous forest maintains a population of earthworms, which are the most important agents in distributing organic matter throughout the layers which they occupy. And thus, whilst the organic matter profile under conifers consists of a sharply defined raw humus layer, overlying a mineral soil relatively poor in organic matter (although sometimes including a humus B horizon), the organic matter of the soil of a deciduous forest shows a fairly uniform distribution throughout the top foot or eighteen inches, decreasing gradually in deeper layers. An accumulation of organic matter in a B horizon such as occurs in the humus podsols is never found in the brown earths.

The marked translocation of sesquioxides, which is be-tokened by the bleaching of the A horizon and the brown or reddish-brown coloration of the B horizon in podsols, is nearly or completely absent in the brown earths, where bleaching is scarcely shown. Apart from mechanical translocation, the clay complex remains comparatively unaltered throughout the profile.

A. Stebbutt considers that, in the brown earths, the clay complex has undergone a certain degree of degradation into its constituent silicic acid and sesquioxides. The brown colour of these soils is an evidence of the presence of free hydrated ferric oxide. A further stage in the degradation process consequent on impoverishment in bases would involve the differential leaching out of the sesquioxides and the development of a bleached A horizon as in the podsols.

Structurally, the brown earths show important differences from the podsols. Whilst in the A horizons of podsols the single grain structure predominates, the brown earths are generally distinguished by a crumb structure or a variant of this structure. This may be attributed to the better base status and to the better distribution of base-saturated organic matter, throughout the soil layers.

It should be added that all gradations are possible from podsols to brown earths, and it is thus possible to describe
certain soils as slightly podsolized brown earths. The brown podsolic soils of the United States would appear to form a transition between brown earths (grey-podolic) and podsolics.

**Sweden.** The character of the brown earth profile is illustrated by the following Swedish profile described by K. Lundblad\(^27\). It was found under beechwood.

- **A\(_0\):** Beech litter, 2-3 cm. thick.
- **A\(_1\):** Loamy layer containing mild humus with indefinite lower limit. Thickness 12-15 cm.
- **B:** Brown earth. Thickness 40 cm.
- **C:** Sandy, stony, moraine, principally of granite material.

**England.** W. M. Davies and G. Owen\(^6\) give the following description of a brown earth in Shropshire:—

- **0-9 in.** Brownish-red, loamy sand. Crumbly and somewhat gravelly.
- **9-18 in.** Similar material, but rather more compact. Gravelly. Below 18 in. Disintegrated lavender coloured sandstone, passing into a layer of purplish-red marl at 40 in.

The principal analytical figures are shown in Table XXX.

**Table XXX.—Analytical data for Shropshire brown earth (Davies and Owen)**

<table>
<thead>
<tr>
<th>Depth</th>
<th>0-9 in.</th>
<th>9-18 in.</th>
<th>Below 18 in.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>pH</strong></td>
<td>...</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td><strong>Clay</strong> (%)</td>
<td>4.77</td>
<td>5.10</td>
<td>5.34</td>
</tr>
<tr>
<td><strong>Ignition loss (%)</strong></td>
<td>16.3</td>
<td>17.5</td>
<td>15.3</td>
</tr>
<tr>
<td><strong>Clay SiO(_2)/Al(_2)O(_3)</strong></td>
<td>4.95</td>
<td>3.28</td>
<td>3.00</td>
</tr>
<tr>
<td><strong>SiO(_2)/R(_2)O(_3)</strong></td>
<td>2.44</td>
<td>2.48</td>
<td>2.54</td>
</tr>
<tr>
<td><strong>Al(_2)O(_3)/Fe(_2)O(_3)</strong></td>
<td>2.06</td>
<td>2.04</td>
<td>2.07</td>
</tr>
</tbody>
</table>

The profile is somewhat more acid than is usual in brown earths. The constancy in composition of the clay fraction should be noted. No figures for organic carbon are given, but the ignition loss suggests low figures.

The profile described above is probably typical, in its essential features, of large areas of similar soils in England.
and Wales, and may be expected in well drained situations with parent rocks of moderate base-status under British climatic conditions. Where, as in certain coarse sands, notably in the Trias, the base reserves are poor, podsolks are developed. It is thus possible for brown earths and podsolks to occur under exactly similar climatic conditions, the difference in the final result being due to the nature of the parent material.

**Wales.** The characters of a Welsh brown earth profile are shown in the following description:—

Llanfairpwl, Anglesey. Altitude, 120 feet. Mean annual rainfall, 38 inches. Mean annual temperature, 50°F. Parent material, Pre-cambrian schistose drift. Artificial grassland, but originally deciduous forest.

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Depth</th>
<th>Description</th>
<th>Clay %</th>
<th>Organic Carbon</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>0-9 in.</td>
<td>Dark brown stony loam</td>
<td>18.7</td>
<td>2.9</td>
<td>5.40</td>
</tr>
<tr>
<td>II</td>
<td>9-15 in.</td>
<td>Brown stony loam</td>
<td>20.0</td>
<td>2.05</td>
<td>5.64</td>
</tr>
<tr>
<td>III</td>
<td>15-36 in.</td>
<td>Reddish-brown stony loam</td>
<td>16.9</td>
<td>1.1</td>
<td>5.94</td>
</tr>
<tr>
<td>IV</td>
<td>36-48 in.</td>
<td>Yellowish-brown stony loam</td>
<td>n.d.</td>
<td>nil</td>
<td>5.80</td>
</tr>
</tbody>
</table>

There is no well marked differentiation between A, B, and C horizons. The variations in the composition of the clay fraction are shown in the following table:—

<table>
<thead>
<tr>
<th>Horizon</th>
<th>SiO₂/Al₂O₃</th>
<th>SiO₂/R₂O₃</th>
<th>Al₂O₃/Fe₂O₃</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>2.39</td>
<td>1.80</td>
<td>3.05</td>
</tr>
<tr>
<td>II</td>
<td>2.47</td>
<td>1.85</td>
<td>2.99</td>
</tr>
<tr>
<td>III</td>
<td>2.41</td>
<td>1.80</td>
<td>2.95</td>
</tr>
<tr>
<td>IV</td>
<td>2.64</td>
<td>2.05</td>
<td>3.47</td>
</tr>
</tbody>
</table>

These figures exhibit a reasonable constancy in composition of the clay fraction in the first three layers. The fourth layer may well represent C horizon material. So far as an A horizon is concerned, either it has been removed by erosion or else it has been mixed by cultivation with the B horizon. Neither the field characters nor the laboratory data offer any evidence of an eluviated horizon.

**N.W. Europe and N. America.** Brown earths occur commonly over a large area of Western Europe. They
Grey-brown podsolic soil, Tennessee.
occur also in the Eastern United States from New York to N. Carolina. The American brown earths, or grey-brown podsolic soils, often show a marked development of a texture profile due to the mechanical eluviation of clay and fine material from the surface horizons. From the data available it would appear that there is no marked difference in the composition of the clay fraction in the different horizons, as would be the case if chemical eluviation had been operative. The removal of clay from the surface horizons does, however, produce a weakening of the soil colour in the top soil, so that many of these soils appear somewhat bleached. It should be added that in many districts there has been extensive sheet erosion, with the result that the original B horizon is now exposed, and in many cases forms the cultivated soil.

J. Geering\textsuperscript{28} has described the development of brown earths on soft limestone (molasse) in Switzerland. Basing his discussion on the composition of the clay fractions in the profile, he proposes the use of a quotient obtained by dividing the clay $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio in the A horizon by the corresponding ratio in the B horizon. In the profiles examined by him this quotient varies from 1.0-1.1, whilst in podsols the corresponding figures are 2.5-4.9.

BRITISH AGRICULTURAL BROWN EARTHS

Most of the soils of Great Britain with good natural drainage were probably brown earths under primitive conditions, but owing to their long agricultural history, they must be regarded as highly artificial. The original deciduous forest that covered most of the country has been almost entirely removed and authentic natural profiles can be studied only in isolated localities such as the Forest of Dean, the New Forest, and the Forest of Wyre. It is to be regretted that, at present, no data are available for profiles in these areas.

Under ordinary farming conditions, the incorporation of organic matter with the soil by the addition of farmyard manure and by the cultivation of artificial grass leys, and the
maintenance of the base-status by chalking, liming, or marling, have tended to modify the original brown earth soil and produce a type of soil which in some respects resembles the prairie soils to be discussed below.

We have seen that brown earths are developed under deciduous forest and that the relatively high mineral content of the leaf litter and the presence of a gramineous and herbaceous ground vegetation maintain the base-status of the soil and oppose podsolization. Where the natural deciduous forest is removed and the soil is afforested with conifers, it may happen that deterioration in base-status proceeds to such a degree as to result in incipient podsolization. This may be seen in plantations on light soils in England, for example, in the Delamere Forest of Cheshire. Similar deterioration takes place where natural deciduous forest becomes replaced by heath. Podsolization is a deteriorative change, and it is important to realize that, on light soil under heavy rainfalls, the change from brown earth to podsol may easily follow the removal of deciduous forest. G. Lemée has described the degeneration by podsolization under Calluna of certain soils in the Perche district of France, formerly under oak forest. The parent material is quartzose sand.

PRAIRIE SOILS

The podsols and their congeners are for the most part developed under forest. Beyond the forest limit to the north lie the tundras (Chapter XII), where, although the humidity is moderate or excessive, the general rigour of the climate precludes forest growth. There is an important group of soils, termed prairie soils, lying on the arid side of the eastern forest belt of the United States and occupying a considerable proportion of the so-called Middle-West. These soils, although developed under steppe or prairie conditions, appear to belong to the great group of humid soils rather than to the arid soils, which lie adjacent to them and are also developed under steppe conditions. To the east of the prairie belt occur the podsols in the north, the brown earths (grey-
brown podsolic soils) in the middle, and the red and yellow podsolic soils (Chapter XIV) in the southern states. The rainfall under which the prairie soils occur is sufficient to produce a complete leaching of soluble salts, including carbonates, in the soil profile, and there is thus no horizon of accumulation of calcium carbonate or gypsum as in the neighbouring chernozems (see Chapter XI). Calcium carbonate encountered in a prairie profile belongs to the C horizon and is to be reckoned with the parent material. A characteristic prairie soil profile consists of three horizons, namely, a dark coloured horizon rich in organic matter, a transition horizon often showing brown colours, and the underlying parent material, which is generally of a grey colour. Calcium carbonate is found at a depth of 50-60 cm. where, as is frequently the case, the parent material is a calcareous loess.

The following profile, developed on loess in Houston Co., Minnesota, is described by A. L. Gray, et al.\textsuperscript{30}.

0-9 in. Very dark greyish-brown mellow silt loam, nearly black when moist. Granular throughout and laminated. Grey specks in upper 5-6 inches, more noticeable in lower 3-4 inches.


19-30 in. Light brown silt loam with large aggregates, tending to blocky structure, slight lamination in upper 1-2 inches. Grey coatings on aggregates. Some organic material on surface of aggregates. "Abandoned worm and insect channels abundant."

30-49 in. Parent material with transitional layer in top 6 inches. Yellowish-brown silt loam with no noticeable structure. Some light grey mottling in lower 8-9 inches.

The granular structure is not so marked as in the chernozems and may be entirely lacking, whilst the calcium carbonate in the C horizon is quite distinct from the concretionary deposits found in the more arid soils.

The prairie soils resemble the chernozems in their organic matter profile. The abundant and deep rooted grass vegetation, the presence of a rich soil fauna, and, in spite of
the leaching of calcium carbonate, the favourable base-status, all result in a deep enrichment of the soil in dark-coloured base-saturated organic matter.

Eluviation of sesquioxides and deposition in a B horizon does not occur. It may be conjectured that the clay complex in prairie soils has scarcely suffered to any appreciable extent that degradation into free silicic acid and sesquioxides which Stebbut postulates for the brown earths. The dark colours of the prairie soils mask field evidence of changes affecting the inorganic complex, but incipient degradation is sometimes shown by the presence of brownish colorations below the dark coloured horizon.

It should be added that H. Stremme\(^{31}\) and certain other continental writers consider the North American prairie soils as "meadow" or "gley" soils (see Chapter XI), whose distinctive characters are due to the presence of groundwater. On the other hand, N. Florov\(^{32}\) assigns these soils to the class of degraded chernozems (see opposite). The American workers class them definitely with the humid soils and relate them to the podsol group. It is considered that the prevalence of steppe rather than forest conditions is due to the prevention of natural regeneration through grazing animals and prairie fires. It is certain that, since the extinction of the herds of bison, and the prevention of fires by settlement, forest vegetation has tended to invade these regions, and it may be reasonably conjectured that, in the absence of interference, the prairie soils would undergo a kind of degradation analogous to that exhibited in the forest soils bordering on the chernozem regions of Russia. (See below.)

Prairie soils doubtless occur in many parts of the world on the humid side of the boundary between the humid and arid soils. Economically, they must rank among the most fertile of the great groups, and their recognition and definition is an important task for the future.

There is a class of soils occurring in the humid west of Britain that may be considered to have some affinities with the prairie soils. These are the soils of permanent grass-
lands. The resemblance consists in their high content of organic matter well distributed throughout the soil profile. But, inasmuch as, except on calcareous soils, their base-status is often low, the type of organic matter formed by micro-biological decompositions will be greatly different, being of a less marked black colour. Further, the clay complex under such conditions is less stable than in the prairie soils and may undergo degradation, as in the brown earths, or even differentiation, if the profile is sufficiently acid to permit podsolization.

DEGRADED CHERNOZEMS

We have seen that the prairie soils of N. America occur in a belt between the forest soils (podsols, etc.), and the chernozems to be described in the next chapter. In Europe their place is occupied by a group of soils termed degraded chernozems or grey forest soils. It is considered that the region occupied by these soils represents former steppe and that the profiles found there have certain surviving characteristics of the chernozems.

In the first stages of degradation under forest, the changes are mainly structural and consist in the destruction of the characteristic porous crumb structure of the chernozem. There is also a sinking of the calcium carbonate horizon and some loss of organic matter. In a more advanced stage of degradation, there is a further obliteration of the porous structure and the development of a mineral profile, characterized by eluviation and deposition of sesquioxides and the disappearance of carbonates from the profile, resulting in the formation of the so-called grey forest soil.

The following data by Georgievski are quoted by Glinka for a grey forest profile from the Poltava government:

\[ A_0 \text{, } 0-2.5 \text{ or } 5 \text{ cm. } \text{Forest litter.} \]

\[ A_1 \text{, Dark brown to light grey granular layer, the granules increasing to walnut size, down to } 26 \text{ cm., where there is a change to} \]
SOILS

A2. Ash grey horizon with nut structure. On drying, breaks up into small angular blocks with greyish-white powdery coating. The thickness of this horizon is 47-48 cm.

B1. Reddish-brown compact loam at first containing organic matter and showing nut structure. Dark brown deposit in pores and cracks. The thickness of this horizon is 70-140 cm.

B2. Brownish, highly calcareous loam sometimes passing into hard white marl. Thickness 70-140 cm.

C. Yellow loess.

A transition occurs to typical podsol soils on the side of increasing humidity.

The decisive feature of the degraded chernozems or grey forest soils is the development of a brown or reddish-brown colour in the subsoil horizons, indicating the operation of podsolic eluviation. The decrease in organic matter content compared with adjacent black earths has been shown by numerous analyses. It appears, however, that the organic matter profile shows a more gradual decrease of this constituent with increasing depth than in the typical podsols.

The Russian workers have recognized a number of stages in the degradation of chernozems, ranging from the unchanged chernozem to definite podsolic forest soil with markedly bleached horizons below the forest litter. H. Stremme\textsuperscript{34} and others have described soils in Germany which appear to belong to the class of degraded chernozems. Examples are given by C. E. Kellogg\textsuperscript{35} from Wisconsin, where grey forest soils appear to have developed from former prairie soils. M. Popov\textsuperscript{36}, in Rumania, has described the successive gradations from chernozem through degraded chernozem to brown-red forest soil in the Dolj district. The parent materials are diluvial clay or loessial loams. The brown-red forest soils are probably akin to the N. American red podsolic soils. Degradation of the chernozem is marked by incipient eluviation and development of an illuvial B horizon. This is accompanied by a sinking of the calcareous horizon and fall of base-status, accentuated in the final brown-red forest soil stage.
SOILS OF THE PODSOLIC GROUP

MOUNTAIN SOILS

The soils of alpine regions resemble soils of high latitudes, principally in the circumstance that, owing to the prevalence of low temperatures, weathering is mainly of a physical character. Apart from the low temperatures under which they are developed, the most important factor affecting them is the strong relief, which facilitates erosion and, except in valleys, prevents the accumulation of deep layers of weathered material.

Generally speaking, there is a high precipitation in mountain regions. This is not invariably the case, and it may also happen that the rainfall does not increase uniformly with altitude, but passes through a maximum at intermediate altitudes. The humid character of mountain regions favours intense erosion, and the soils on hillsides are generally thin, whilst valleys, if not occupied by glaciers, are deeply filled with the spoil of the higher lands. The intense leaching results in removal of bases so that soils are generally acid in character. H. Jenny found that, even on calcareous rocks, about 80% of alpine soils examined in Switzerland were actually acid. The acid character of the soil and the low temperatures which prevail favour the accumulation of peaty organic matter, the so-called alpine humus, in folds and hollows. Generally, however, erosion prevents the accumulation of more than about 4 inches of rocky humus soil.

The relief is generally the most important factor in differentiating soils of mountain regions. Thus, whilst in exposed situations the soil may be represented by bare rock or spreads of angular rock fragments, on gentle slopes and shelves there may be a sufficient depth of soil to carry a closed association of grass or, at lower elevations, forest.

Alpine humus soils occur above the level of forest. Whilst humidity generally increases with altitude, at great elevations, maximum humidity is passed and a decrease sets in. This decrease may in some cases determine the recession of forest growth and the prevalence of grass vegetation. In other cases, the limit may be imposed by the rigour of the
climatic conditions. The height at which forest soil or its successors gives place to alpine humus varies with latitude. In Britain the limit may lie at about 2,000 feet, whilst in equatorial regions it may be as high as 13,000-14,000 feet.

In a study of the soils of a mountain area in N. Wales, G. W. Robinson and T. Waśowicz²⁷ found podsols and humus podsols up to nearly 3,000 feet, with alpine soils on the highest peaks. Truncated soils were also found with surface soils of strongly marked B character.

The profile of alpine humus soils is relatively simple and consists of a layer of humus soil overlaying rock. The thickness is variable but generally small. The organic matter content varies considerably. Whilst in some cases the soil may consist entirely of a felt of fibrous peaty matter, in other cases only a few per cent. of organic matter is present.

The vegetation of alpine humus soils is of a decidedly xerophilous character. The conditions, however, differ markedly from those obtaining in deserts, for the adaptation is not to predominantly arid conditions but to strongly variable conditions, which include periods of intense drought. The drought is intensified by the thinness of the soil, the prevalence of high winds, and the strong insolation. Xerophytic characters can be observed also in heath vegetation at lower altitudes.

The soils in wet hollows at high altitudes are naturally of a peaty character. Thick layers of peat may be developed where slopes are maintained constantly wet by surface water.

Alpine soils are generally dark in colour owing to the presence of organic matter. The colour of the mineral soil is generally grey. Owing to the low temperatures, there is no chemical weathering and thus no formation of colloidal clay to undergo degradation as in the podsols. Whilst clay of primary origin is not likely to occur in high alpine regions, it may happen that, where the parent rock consists of unmetamorphosed sediments, soils containing colloidal clay will occur and podsolization may be expected.
Owing to their inaccessibility, comparatively little is known as to the conditions of soil formation at high altitudes. The altitude at which the distinctive characters of mountain soils occur varies, of course, with latitude, and whilst in Central Africa these characters are seen only near the peaks of the highest mountains, in high latitudes they are encountered at comparatively low elevations. Indeed, it might be thought that the principal distinction between alpine soils and the tundra soils of high latitudes consists only in the nature of the relief forms. The important difference consists in the much more intense solar radiation in equatorial regions. The exact consequences of this remain to be elucidated.

PODSOLIC SOILS WITH HIGH BASE STATUS

It is customary to think of podsols as developed under extremely acid conditions and indeed to regard acidity, i.e., a relatively high hydrogen-ion concentration as a condition of their formation. It is possible, however, as suggested by P. H. Gallagher, that the effective agent in podsolization is not the hydrogen-ion but the anions of organic acids, formed by the decomposition of plant residues. P. H. Gallagher and T. Walsh have described podsolic soils, termed by them grey-brown soils, developed in calcareous drift in Co. Kildare, Eire. Whilst the grey horizons and the underlying brown horizons are suggestive of podsols, the reaction of the solum is on the alkaline side of neutrality. The characteristic superficial layer of raw humus is also lacking. Yet the ratios for the clay fractions confirm the impression obtained from the field characters that there has been a translocation of sesquioxides from the A to the B horizon. A similar profile is described from Co. Roscommon.

In many parts of N. America there are soils under mixed deciduous and coniferous tree cover which resemble podsols in morphology but are developed in calcareous drifts. These soils, sometimes termed "grey wooded soils," were discussed by A. Leahey and by H. C. Moss, who preferred to assign them to a separate group. In this connexion, it is
of interest to note that H. M. Galloway has reported an alkaline raw humus from Wisconsin, but as it lies directly on shattered limestone no podsol is observable.

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1 Georgievski; from K. D. Glinka, Die Typen der Bodenbildung, Berlin, 1914, p. 68.
17 Lebedev, A. F., and Bankova, E. E., Physical characteristics of the soil profile, Moscow, 1930.
SOILS OF THE PODSOIL GROUP


23 Mattson, S., and Gustafsson, Y. See reference 5.


33 Georgievski; from K. D. GLINKA'S Die Typen der Bodenbildung, Berlin, 1914, p. 90.


CHAPTER XI

CHERNOZEMS AND THEIR RELATED GROUPS

The soils to be described in the present chapter are developed in semi-arid and arid climates with incomplete leaching, which results in the development of horizons of calcium carbonate deposition. They are developed under steppe or desert vegetation. The term podocal, suggested by Marbut, may be conveniently used to cover the groups comprised in the succession from black earths to grey-brown and desert soils.

CHERNOZEMS* OR BLACK EARTHS

The study of the chernozems or black earths is of considerable interest and importance, not only on account of their wide extent and agricultural value, but because the modern study of soils in their natural relationships received its greatest impetus in the investigations and discussions of the problems of the origin of these and related soils. Further, regarding the diversity of the soils of the world as resulting from the varying intensity with which the pedogenic processes depending on temperature and moisture have operated, it is apparent that the black earths occupy, as it were, a central position between extremes with respect to the intensity of the factors that have affected their development.

The obvious feature of the black earths is, as the Russian name implies, their colour, which is due to the presence of humified organic matter. In the centre of the

*This spelling is used by the Commonwealth Bureau of Soil Science and by the U.S. Bureau of Soils, and I have therefore adopted it in place of "tshernosem," used in former editions.
chernozem area in Russia, where the conditions are most
typical for the development of these soils, the colour is deep
black, but as the adjoining soil provinces are approached,
the black colour becomes less marked, changing into the
grey of the degraded chernozems on the one side and the
brown of the chestnut earths on the other.

The intensity of the dark colour produced would lead an
observer familiar only with soils of humid climates to expect
larger proportions of organic matter in the chernozems than
actually occur. A sample of chernozem from the Ukraine,
examined by the writer, contained only 3.2% of organic car-
bon, i.e., about 6% organic matter; yet the dark colour sug-
gested a much higher proportion. As much as 16% of
organic matter may be present in the central chernozems,
but 8 to 10% is more usual. The dark colour may be attri-
buted to some extent to the fact that these soils are base-
saturated, and that the humification processes proceed in a
neutral or even alkaline medium. It would seem, however,
that other factors must be involved. Certain Lower Lias
soils in Britain contain free calcium carbonate and notable
proportions of organic matter, yet do not exhibit the dark
colour of the chernozems. In all probability the high tem-
perature to which these soils are subjected during summer,
in conjunction with the high base-status, may determine
their dark colour. It is noteworthy that, in many parts of
the tropics and sub-tropics, dark coloured soils are asso-
ciated with a high base-status, whilst base-unsaturated soils,
under similar climatic conditions, show little or no humic
colour, although they may actually contain higher propor-
tions of organic matter than neighbouring dark coloured
soils.

The thickness of the dark coloured layer in the cherno-
zems varies considerably. The average for the best
developed black earths is about 70 to 100 cm., with 40 cm.
to 150 cm. as extremes. It is considered to be differentiated
into two horizons, an upper, A₁, and a lower, A₂, the latter
having a lighter colour and a lower organic content. The
thickness of the dark horizons and the organic matter content decrease towards the boundaries of the soil group. The sharpness of the transition to the underlying light-coloured parent material varies somewhat, but the change generally takes place within a few centimetres.

With regard to the inorganic profile, the typical chernozem shows complete leaching of soluble salts, incomplete leaching of calcium carbonate and calcium sulphate, and no leaching of silica or sesquioxides. Thus, whilst there is no accumulation of sodium salts in the profile, an unfailing mark of the chernozem is the zone of accumulation of calcium carbonate, sometimes accompanied by calcium sulphate. The absence of soluble salts in these soils of incomplete leaching may possibly be due to the fact that occasional abnormally wet seasons provide an excess of rainfall over evaporation, so that the profiles are completely leached. Such occasional leaching may be sufficient to remove any soluble salts that would otherwise accumulate in the profile.

The depth of this zone will vary according to the parent material. It approaches the surface more nearly the more calcareous is the parent material and also the more arid is the climate. The calcium carbonate zone may occur in the humic layer, but is generally near or below its lower limit. Above the main calcium carbonate layer, calcium carbonate occurs deposited in fibrous threads, like fungal hyphae, and in small concretions which are sometimes hollow and may assume curious forms. Gypseous deposits may occur below the calcium carbonate horizon. In some cases, a second horizon of calcium carbonate accumulation may be observed at a depth of a few metres, but this would not appear to be a necessary characteristic.

A characteristic peculiarity of chernozems, and indeed of the adjacent chestnut coloured earths, is the presence of *krotovinas*. These are formed by the filling-in by calcium carbonate of the burrows of steppe animals.

The black earths are characterized by their well-developed granular crumb structure in the humic layer.
This is, of course, associated with their high base-status and the chernozem is thus the chief example of a "calcium soil." Although these soils may contain high proportions of clay, they offer few difficulties in tillage and consistently maintain their crumb structure.

So far as the clay complex of the soil is concerned, there would appear to be no vertical differentiation. It is unfortunate that no analyses of the clay fraction of Russian black earth profiles are available; but there is no evidence from field observations or from laboratory data to suggest the eluviation of sesquioxides or silicic acid. Neither is there any evidence of the eluviation of the clay complex as such, with consequent development of a texture profile. It is held by A. Stebutt that, in the chernozems, the conditions for the formation and maintenance of the aluminosilicic weathering complex are at optimum. Increase of temperature and increase in humidity both favour the destructive processes whereby the component silicic acid and sesquioxides are liberated and, in the podsols, differentiated.

The brown or reddish-brown colours associated with the presence of free sesquioxides are entirely lacking in the typical chernozem profile, in which the mineral colour is predominantly brownish-grey. The transition to degraded chernozems in the direction of increased humidity is marked by the appearance of brown sub-soil horizons. The transition to higher temperature conditions is marked also by the development of reddish-brown colours throughout the profile, indicating the destruction of the weathering complex without eluviation.

In Russia, the black earth region is characterized by cold winters and hot summers, with a moderate rainfall. Data presented by P. Kossowitsch² show mean January temperatures from —20.8°C in Siberia to —3.5°C in the Ukraine, and mean July temperatures varying from 16.7°C in Siberia to 23.3°C in S. Russia. The mean annual rainfall varies from 300 mm. to over 500 mm. A considerable proportion of the precipitation is in the form of snow, and the
summer rainfall is heavily outweighed by evaporation. These conditions result in steppe vegetation which grows luxuriantly in spring and in early summer. The ensuing drought puts an end not only to active plant growth but also to micro-biological activity in the soil.

C. C. Nikiforoff has discussed the general problem of chernozem formation. Concerning humification, he points out that each season's growth is returned as residues under natural conditions. The mat of dead vegetation of each year begins to decompose in the following spring and the process extends over succeeding years, so that each cycle overlaps phases of preceding cycles. This author considers that the pedological origin of the calcium carbonate accumulation has not been definitely proved. Most chernozems have in fact been developed from lime-rich parent materials.

In Russia and Asia, where the black earths have been principally studied, the parent material is generally loess and the textural range is limited by the character of this deposit. The extent of the loess and of the chernozems are, however, not necessarily identical, for not only do other soil groups occur on the loess, but chernozems may also be developed from other parent materials.

K. D. Glinka describes the following profile from Kogalnitza, Rostov:

A₁. 0-20 cm. Black with brownish-grey shading, breaking down into irregular lumps and eventually into a porous fine grained material. Effervescence with HCl from 8 cm.
A₁. 20-55 cm. Greyish-black granular; more cohesive below. Columnar and nut structure.
A₂-C. 105-185 cm. Yellowish-brown clay with vertical tongues of humous material. Calcium carbonate concretions.
A₂-C. 185-240 cm. Humus streaks in wet pores. Decrease in carbonate concretions.
C. Yellowish-brown clay.
Germnay. Soils which appear to belong to the black earths occur in many parts of Germany. The following profile is described by K. Von See, from near Magdeburg:—

A. 0-55 cm. Humous carbonate-free loess. Loose and porous. Below the root horizon down to 30 cm. a greyish-brown horizon with lump structure over an almost black humous layer with tendency to polyhedral structure. Occasional flecks of greyish-yellow carbonate-free loess. At the transition to parent loess a zone rich in carbonate.

C. Yellow or yellowish-grey unstratified loess.

V. Hohenstein describes numerous chernozem profiles developed on boulder marls, clay marls, and glacial sands, in Eastern Germany. They are characterized by the leaching of carbonate from the upper horizons down to 20-100 cm. and the development of a zone of carbonate accumulation. The humus content, 2-4%, is rather less than in the Russian black earth region. The prevailing colours are coffee-brown to dark brown. They are considered to have been formed under an earlier steppe climate.

W. Laatsch has discussed the distribution and development of soils of the chernozem type developed on loess in the vicinity of Halle. It appears that the conditions in this region represent the limit of stability for such soils, for there is a tendency towards degradation and the production of brown forest soils. Where the mean annual rainfall is greater than 500 mm. the brown forest soils are developed on loess without an intermediate chernozem stage.

Czechoslovakia. V. Novák describes "Hanna" soils in Czechoslovakia which appear to belong to the chernozem group. A typical profile from Hulin showed:—

0-40 cm. Dark loam with brownish tinge. Loose crumb structure sharply delimited below.
40-70 cm. Compact humous loam easily breaking down into crumbs.
70-100 cm. Dark compact loam.
100-150 cm. Yellowish loam.
150-350 cm. Light coloured loam with concretions and superficial deposits of calcium carbonate.
The mechanical texture is very uniform. The humus content decreases gradually from 4.83% in the surface horizon to 3.67% at 100 cm. There is then a sharp fall to 2.00% at 120-140 cm.

Another type of Hanna soil differs from that described above in being lighter in colour and poorer in humus. Calcium carbonate concretions appear at 130-140 cm.

North America. Chernozems occur over considerable areas of North America in the States or Provinces of Saskatchewan, Alberta, Minnesota, the Dakotas, Nebraska, Kansas, Colorado, Oklahoma and Texas. Whilst the northern soils resemble the Russian chernozem, some change in character occurs in passing to the warmer conditions of lower latitudes.

A northern chernozem from N. Dakota, mapped as the Barnes series (see Plate VI), is thus described by C. F. Marbut:

I. 0-12 in. Heavy black loam. Irregular clod structure in top 4 in., ill-defined columnar structure below.

II. 12-22 in. Dark greyish-brown clay loam with ill-defined columnar structure. No effervescence with acid.

III. 22-32 in. Light greyish-yellow to greyish-brown, heavy silt loam or silty clay loam. White spots and splotches of calcium carbonate. Columnar structure disappears in this horizon, which is marked by maximum of carbonate accumulation.

IV. 32-40 in. Mottled silty clay loam. Small iron oxide stains and concretions present. Spots and concretions of calcium carbonate less abundant than in III.

V. Greyish-brown silty clay loam with iron oxide stains and concretions. Calcium carbonate present in disseminated form. This is the parent glacial drift from which the profile is developed.

The following is Marbut’s description of a southern chernozem from Texas, mapped as the Amarillo series:

CHEKNOZEM PROFILE, CASS COUNTY, N. DAKOTA, U.S.A.
II. Dark brown to very dark brown, generally clay loam. Dark yellowish-brown or reddish-yellow on being crushed. Granular structure becoming coarser with depth. Thickness about 12 in.

III. Reddish-yellow clay loam, still with darker coloured coatings in structural elements. General tendency to columnar structure.

IV. Reddish-yellow material with much calcium carbonate. Similar structure to III.

V. Parent material; unconsolidated calcareous silt and clay.

Data from M. S. Anderson and H. G. Byers\(^{10}\) for a Texas profile, which, though somewhat different from the Russian chernozem, is essentially similar in its general type of development, are given in Table XXXI. The soil series is the same as that described above.

<table>
<thead>
<tr>
<th>Depth</th>
<th>SiO(_2) (%)</th>
<th>Al(_2)O(_3) (%)</th>
<th>Fe(_2)O(_3) (%)</th>
<th>SiO(_2)/R(_2)O(_3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-5 ins.</td>
<td>50.51</td>
<td>22.04</td>
<td>8.80</td>
<td>3.10</td>
</tr>
<tr>
<td>10-20 ins.</td>
<td>51.51</td>
<td>22.71</td>
<td>8.61</td>
<td>3.09</td>
</tr>
<tr>
<td>30-40 ins.</td>
<td>51.32</td>
<td>22.43</td>
<td>8.46</td>
<td>3.13</td>
</tr>
<tr>
<td>54-64 ins.</td>
<td>51.23</td>
<td>24.08</td>
<td>8.19</td>
<td>2.97</td>
</tr>
<tr>
<td>70-75 ins.</td>
<td>38.42</td>
<td>17.64</td>
<td>5.71</td>
<td>3.06 (CaCO(_3) horizon)</td>
</tr>
<tr>
<td>96-100 ins.</td>
<td>45.88</td>
<td>20.09</td>
<td>6.89</td>
<td>3.18</td>
</tr>
</tbody>
</table>

The composition of the clay colloid shows a remarkable constancy throughout the profile and there is no evidence for eluviation either of silicic acid or sesquioxides.

H. W. Hawker\(^{11}\) describes a range of soils in the Rio Grande valley of Texas which show successive stages of leaching from young delta soils to mature terrace soils with calcium carbonate accumulations at 4 to 8 feet. The mean annual rainfall is 22.5 inches and the profile consists of light brownish-grey to greyish-brown soil overlying heavy bluish-grey clay, mottled yellow and, sometimes, red. The surface horizons are non-calcareous.
India. There is in India a well known type of soil known as the regur, or black cotton soil, which occupies a large area in the southern half of the Peninsula. The natural vegetation appears to be of a grass steppe type. Regur is usually of a clay character and varies in depth from 1 to 2 metres, occasionally reaching 5 metres. On drying it shrinks and cracks to such an extent that it is often said that “black soil ploughs itself.” Like the chernozem, it is characterized by a zone of calcium carbonate concretions (“kunkur”). The parent material is generally a black basalt trap, the so-called Deccan trap. The organic matter content is generally low and the black colour, according to H. E. Annett, is due in part to the presence of finely divided titaniferous magnetite.

A. D. Desai has described black cotton soils of the Hyderabad-Deccan state in India. These soils are generally heavier in texture than the adjacent red soils. They are derived from the Deccan trap and gneissic complex. The depth and development of the calcium carbonate horizon varies considerably and may be as much as 8 to 10 feet and, in exceptional cases, 30 to 40 feet.

The climatic conditions for the Indian regur soils differ from those in the Russian chernozem region in that the temperature, even in the winter months, is always considerably higher. The rainfall, rather more abundant than in the Russian black earth region, is, however, insufficient to compensate for the great evaporation resulting from the high temperatures which prevail, and incomplete leaching results. The climate is thus not identical with that of the Russian chernozem region. Regur might, therefore, be described as tropical black earth. Black soils occur in Morocco, Kenya, and other parts of Africa, which may have affinities with the black earths, and are, probably from their resemblance to regur, also termed “cotton soils.”

*W. H. Harrison and M. R. R. Sivan (Mem. Dept. Agr. Ind. Chem. Series, 1912, II. No. 5) only admit Annett’s explanation for the trap areas. Elsewhere, colloidal iron and aluminium silicates are held responsible for the black colour.
Argentina. The soils of the Argentine Pampas would appear also to present many similarities with the black earths and further work may result in their being assigned to this group.

The black earth represents the soil type developed under the conditions which prevail on the arid side of the boundary between humid and arid soils. When they have been completely investigated in all parts of the world where they occur, it will probably be possible, as Marbut suggests, to divide them into sub-groups on the basis of the prevailing temperatures. We may thus speak of high latitude, mid-latitude, sub-tropical, and tropical chernozems. The common characteristics are a dark A horizon, relatively rich in base-saturated organic matter, a high base-status, and the presence of an horizon of calcium carbonate accumulation. The vegetation is a closed association of steppe, savannah, or pampas character, either devoid of trees or else with isolated individuals as in the S. African bush-veldt.

BLACK COTTON SOILS OF KENYA

Grey and black soils occur in Kenya which may represent tropical chernozems. D. S. Gracie describes a group of soils, locally termed "black cotton soils," which appear to exhibit various stages of degradation. In the undegraded type, the profile consists of a black clay, rich in exchangeable calcium, overlying a layer of calcium carbonate accumulation. In the degraded soil, there is a grey layer at the surface with acid reaction and low exchangeable calcium content. This overlies the black clay. A still further stage of degradation is seen in certain soils with acid greyish-brown surface horizons. Degradation appears to involve a loss of organic matter. Black cotton soils may also be developed on lighter textured parent materials. They may be analogous with the "vlei" soils (see p. 283) of the humid regions of Kenya, and tend to occur in plains and depressions in relatively hot and dry regions. Their complete
analogy with the chernozems of higher latitudes would appear to be still unproven.

H. H. Bennett and R. V. Allison\^{16} report soils with zones of calcium carbonate accumulation in the drier parts of Cuba, but doubt their analogy with the chernozems of the United States.

CHESTNUT-COLOURED SOILS

The group known as the chestnut-coloured soils (also called chestnut earths) occurs in Russia and Asia on the arid side of the region of the related chernozems. They differ from chernozems principally in the lower proportions of organic matter in the humous layer and in the closer approach to the surface of the zone of carbonate accumulation. Both circumstances are intelligible consequences of the drier character of the climate as compared with that of the chernozem zone.

According to K. D. Glinka\^{16}, the A\textsubscript{1} horizon consists of an upper portion (5-7 cm.) of light colour and loose stratified structure, and a lower portion of smaller thickness and compact structure, lacking the characteristic granular character of the chernozems. The A\textsubscript{2} horizon, which may reach 60 cm. in thickness, is lighter in colour and compact. Both horizons show a prismatic structure by reason of the occurrence of vertical cracks. The organic matter content of the humous horizons varies from 3 to 5%. The mechanical composition is fairly uniform and there is probably no vertical differentiation of the constituents of the clay complex. Accumulation of calcium carbonate may occur in the A\textsubscript{1} horizon or even near the surface. Gypsum may occur in the lower part of the A\textsubscript{2} horizon.

The colours of the humous horizons are greyish-brown and do not appear to be influenced by the presence of free hydrated ferric oxide.

The natural vegetation of the chestnut earths is low-grass steppe.
BROWN AND GREY SOILS OF THE SEMI-DESERT

With passage to more arid conditions, the chestnut-coloured soils are replaced successively by brown and grey semi-desert soils. These differ from the chestnut coloured soils by their lower content of organic matter and by the closer approach to the surface of the calcareous and gypseous horizons. The grey soils are sometimes called sierozem.

United States. In a study of soil zones in northern Wyoming, J. Thorp has been able to trace a complete succession from grey-brown desert soils to podsols in passing from the arid plains to the relatively humid mountain regions. The grey-brown desert soils, developed under mean annual rainfalls of 5-10 inches, show the following profile characters:

A1. 0-1 in. Pale grey-brown "crust and mulch" with fine vesicular structure in dry condition.

A2. 1-6 in. Light greyish to yellowish-brown slightly laminated soil of medium texture.

B1. 6-9 in. Compact cloddy, yellowish-brown soil of slightly heavier texture than above horizon. Most compact horizon in profile.

B2. 9-(18-40) in. Horizon of calcium carbonate accumulation of varying thickness. Calcium carbonate decreasing in lower part of horizon. White or pinkish gypseum often interspersed with upper layers of parent material below carbonate horizon.


A distinct columnar structure is seen in the lower A and B horizons. Calcium carbonate is present up to the surface. In slight depressions the accumulation of calcium carbonate and gypseum becomes more pronounced.

The vegetative cover is scanty and consists mainly of sage brush, Artemisia tridentata and salt sage, Eruotia lanata.

In many parts of the arid regions of the Western United States, the calcium carbonate horizon attains a considerable degree of induration. Under the local name of "caliche," it is frequently used as a surface for roads.
Owing to the incompleteness of leaching in soils of the pedocal class, there is a general tendency for sodium salts to occur in the profile. This is more pronounced in the more arid regions. Where, owing to topography, a ground-water level occurs near the surface, saline or alkaline soils may occur. Such soils may be considered to be analogous to the meadow and peat soils of more humid regions.

Sudan. The cotton soils of the Eastern Gezira in Sudan appear to be grey-brown semi-desert soils of the pedocal class. They are developed from wind-borne material of rather heavy texture. The rainfall is about 400 mm. per annum, of which about two-thirds falls in July and August. The natural vegetation is sparse thorn scrub.

A typical profile, described by H. Greene, shows a surface soil of rather dark brown colour down to 2 feet. Below this is a grey layer which is penetrated by tongues of the top brown soil. In the upper part of the grey layer, ill-defined, detached lumps of grey soil occur surrounded by brown soil. Those nearer the surface contain small specks of calcium carbonate, whilst those lower down contain gypsum. At the lower limit of the grey layer, about four feet, white crumbly aggregates of calcium carbonate occur, amounting, at their greatest concentration, to 3% of the soil. Below this zone are crystals of gypsum in a matrix of yellow-brown soil.

The drying-out which occurs in the dry months of the year leads to the development of numerous cracks which extend for a considerable depth into the subsoil. These are filled with loose material blown from the surface by wind, and there is, in this way, a certain amount of circulation within the profile.

The organic matter content is low—generally below 1%, even in the dark-brown surface soil. In addition to gypsum, sodium salts are present and are closely correlated in amount with the gypsum content.
DESSERT SOILS

These form a rather ill-defined group of soils which have not yet been sufficiently studied in detail. It is generally considered that weathering in the desert is mainly physical, but it must be remembered that even in desert regions rain does fall occasionally and that the subsurface layers may be from time to time sufficiently moistened for hydrolytic decomposition of silicates to take place. Further, intermittent flooding from adjacent upland regions may provide seasonal moisture. Such intermittent supplies of moisture would not affect the actual surface horizons to any great extent, and here the weathering remains substantially physical in character.

It is somewhat surprising that in desert regions evidences of water action are often very spectacular. Although the rainfall is low, it is often intense and with the absence of vegetative cover flood water soon gathers, particularly when streams are fed from wetter mountain regions. And thus in the broad desert valleys extensive alluvial fans and spreads of gravel are common.

Vegetation is absent except for isolated desert shrubs and small trees, such as the cacti and sagebrush of the western American deserts. Such soil development as occurs is unaffected by the processes of humus formation and desert soils are practically humus-free. Wind erosion is very marked, and a common phenomenon is the so-called "desert pavement," consisting of coarse stones or rock fragments from which the originally accompanying fine material has been blown away. Beneath the desert pavement which acts as a protection, fine material is found. The surface of the stones and rocks of the pavement is often covered with a thin brownish film, known as desert-varnish.

C. C. Nikiforoff has studied the pans and crusts that are characteristic of the sub-surface horizons of desert soils particularly in Nevada and southern California. A distinction is drawn between crusts and pans. Crusts are considered to be formed by deposition from ascending solutions during
periods of desiccation. The cementing materials are most commonly calcium carbonate or gypsum, but ferruginous crusts may also occur. Such crusts should not, according to Nikiforoff, be confused with the calcium carbonate horizons deposited in pedocalic soils under higher rainfall and under the influence of a cover of steppe vegetation. Pans, which consist of an accumulation of clay and fine material, are considered to be the result of hydrolytic decomposition of minerals in the intermittently moist sub-surface horizons.

I. C. Brown and M. Drosdoff\textsuperscript{21} describe a profile from the gentle lower slope of a low granitic hill in the Mojave desert. It shows two inches of loose gritty greyish-brown loam, underlain by a reddish-brown gritty clay, which extends down to about fourteen inches, where it grades into the weathered granitic rock traversed by seams of calcium carbonate. On an alluvial fan the surface twelve inches was light brownish-grey structureless gritty sand, underlain by about eighteen inches of reddish brown gritty clay loam with prismatic structure, grading into weakly lime-cemented gritty sand. At three feet there was a sharp change to friable gritty brown sand of granitic origin. The analytical results support Nikiforoff's view that the clay horizon is of hydrolytic and not of illuvial origin.

\textit{California.} In the desert soils of southern California, developed in colluvial fans of material of igneous origin, well-marked pans are found. The degree of development of the pan is made by C. F. Shaw\textsuperscript{20} the basis of the separation of these soils into series. In the mature San Joaquin series, the profile consists of reddish-brown heavy textured soils passing through brown or reddish-brown subsoils to brown or red "iron" hardpan at 40-90 cm., with more or less calcium carbonate, overlying brown pervious unweathered material.

The profiles of San Joaquin and related soils show abundant evidence of chemical weathering, whose occurrence, in a region of scanty rainfall, may be due to natural
irrigation by flood and seepage water from the adjacent mountains.

*S. Africa.* Red and brown soils with calcium carbonate and ferruginous pans are found in the semi-desert Karroo of South Africa.

**POSITION OF EQUATORIAL PEDOCALS**

Pedocalic soils extend over a wide range, being bounded in the north by sub-arctic tundra and extending into equatorial regions. Certain workers have distinguished northern, mid-latitude and tropical pedocals. F. E. Kenchington makes an interesting suggestion for distinguishing two great groups of chernozem-like soils. On either side of the equatorial humid belt there is a desert belt of varying width. To the north of this belt, in the northern hemisphere, the zone of transition to the temperate humid regions is occupied by the classical chernozems and their congeners. To the south, in the zone of transition to the equatorial humid belt, occur a group of pedocalic soils for which the Sudanese term *teen-suda* is proposed. A similar distribution is envisaged for the southern hemisphere. Kenchington would regard the regur soils of India and the pedocalic soils of the Sudan and surrounding regions as belonging to the teen-suda group. In the writer’s opinion this suggestion is a valuable contribution to the study of the pedocals and merits careful examination.

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CHAPTER XII

HYDROMORPHIC SOILS AND PEATS

The term hydromorphic (or hydromorphous) proposed by S. S. Neustreue is a convenient name for all those soils in which profile development is dominated by the effect of sluggish or impeded drainage. Since the presence of excessive moisture implies diminished aeration, it might be permissible to describe these soils as soils of anaerobic metabolism in contradistinction to soils of free aeration, described by H. Del Villar as soils of free aerobic metabolism (suelos de metabolismo libre en medio aereo). In the strict sense of the term we should include with hydromorphic soils those soils in which sodium salts are present or have affected an earlier stage of development. Such soils, however, are more conveniently discussed separately and are, therefore, not included in the present chapter.

The excessive wetness in the profile may arise in two distinct ways. On the one hand, it may be due to the presence of a regional water-table sufficiently near the surface to affect profile development, i.e., when the soil horizons are in the so-called "capillary fringe" or the wetness may be a consequence of the impervious character of the soil material, more particularly in the sub-soil horizons. The impervious character may be inherited from the parent material, as in heavy clay soils, or it may be the consequence of the development of an impervious horizon by pedogenesis as in planosols and gley podsols.

The degree of wetness will be governed to a large extent by the climate. In a wet climate, such as Wales, soils showing impeded drainage are much more frequent than in a drier climate, because there is so much more moisture to dis-
In some climates the wetness may be seasonal as in many tropical soils that are swamps in the rains, but almost completely dry in the dry seasons.

In any given region the distinction between soils of free and soils of impeded drainage is of the greatest importance both for its significance in profile development and for its bearing on the actual behaviour of the soil under agriculture or forestry. Many writers, following N. M. Sibertzev, regard these soils as *intrazonal*, implying that they are in a sense of accidental or local occurrence among the great world groups. Yet, in view of their wide extent, it would appear better to regard them as one of the major soil categories.

**TUNDRA**

These soils occur in sub-arctic regions such as Northern Europe and Asia, and Northern Canada. They lie outside the limits of forest and are occupied by mosses, carices, dwarf willows, festucas, etc. In the tundra region, the dominant factors in soil formation are physical weathering and waterlogging. The latter is occasioned by the scanty evaporation, which results in the prevalence of a humid climate even where the rainfall is comparatively small. Chemical weathering and micro-biological activity are naturally less pronounced than in warmer climates, but are not entirely suppressed. These soils are commonly assigned to the podsolic group, but the prevalence of waterlogging among the conditions of their development suggests that they would be more appropriately grouped with the ground-water soils.

The characteristic feature of the tundra profile is the occurrence, below a certain depth, of a perpetually frozen layer, whose presence is, in some districts, not inconsistent with a fair level of fertility during the growing season.

A typical tundra profile has been described by Ssukatschev. The succession is as follows:

1. Humous grey-brown horizon. Thickness about 3 cm., containing occasional undecomposed plant remains.
II. Yellowish-grey horizon; occasionally greyish-brown, ochreous friable loam. Thickness 3-5 cm.

III. Greyish-blue uniform sticky loam of semi-fluid consistency. Thickness 8-10 cm.

IV. Brownish-yellow loam similar to horizon II., but more compact. Thickness 2-3 cm.

V. A compact brownish-grey solid loamy horizon, often showing at 40-60 cm. dark fluid mottlings apparently of a humous character. Frozen below 79 cm. from surface, but retaining the same character for a further 10 cm.—the limit of exploration.

The bluish-grey horizon is apparently the characteristic mark of tundra, becoming thicker with increase and thinner with decrease of moisture. In sandy profiles, it disappears, presumably owing to better aeration.

When the surface of tundra becomes frozen, the semi-fluid greyish-blue horizon is held under pressure, owing to the expansion of water below 4°C, and this leads, in some cases to an eruption through the thin frozen layer, in other cases, to the surface becoming thrown into hummocks. The eruption of material leads to a mixing of the profile and tends to obliteration of horizons.

Tundra soils result mainly from physical weathering and are consequently poor in clay. They exhibit a natural tendency to peat formation. The peat horizon is often dissected by running water into a characteristic hummocky formation.

The occurrence of a perpetually frozen subsoil in the tundra region has been discussed by C. G. Nikiforoff. Whilst the prevalence of a mean annual temperature below freezing point might sufficiently account for the existence of a permanently frozen layer, it has been held by some investigators that this phenomenon is a survival from the Glacial Period. Perpetually frozen subsoils occur outside the tundra region and may sometimes underlie the more northern podsols, black earths, and chestnut earths. R. R. McKibbin records such an occurrence in northern Manitoba.
MEADOW OR GLEY SOILS AND THEIR RELATED TYPES

The soils described under this heading are termed "Wiesenböden" and "Gleyböden" by continental writers. In the absence of a more suitable term, the name meadow soil or gley soil is used to describe mineral soils whose profile characters are dominated by the occurrence of a high water-table or an impervious layer or layers, impeding percolation.

These soils, which have certain affinities with the podsoils and occur under similar climatic conditions, owe their distinctive characters to lack of aeration through the presence of a high water-table or the restriction of drainage by an impervious layer. The simplest case is that in which a water-table occurs in a soil developed in material permitting free percolation. Here the water-table fluctuates according to the general balance of rainfall and evaporation, being highest in winter and lowest in summer. There are thus three zones in the profile, namely, (1) a surface zone sometimes of negligible depth in which conditions are generally aerobic; (2) the next zone marked by the limits of fluctuation of the water-table in which aerobic and anaerobic conditions alternate; and (3) a zone below the lowest level of the water-table in which conditions are permanently anaerobic and therefore reducing.

The surface horizon in which most of the root development of the natural vegetation takes place is generally dark grey or brownish-grey with a fairly high organic matter content. The greyness may be due to a lateral leaching of iron, for the seepage water in ditches frequently shows deposits of hydrated ferric oxide (ochre). Rusty markings can often be seen along root channels in this horizon.

The second horizon is lighter in colour because of its lower organic matter status, and there is often a sharp transition marking the limitation of root development and humus addition imposed by the proximity of the water-table. The ground colour is grey but there are also rusty streaks, mottlings, or even concretions, resulting from the oxidation
during the aerobic stage of soluble ferrous compounds formed during the anaerobic stage. The decomposition of hydrated ferric oxide in this horizon may attain considerable proportions giving rise to "lake ore" or "bog iron-ore." Under similar conditions in the tropics lateritic horizons are developed (cf. pp. 412 et seq.). Other secondary depositions may occur in this horizon including manganese dioxide, gypsum, and calcium carbonate. There may also be an enrichment in clay, not necessarily through eluviation, but as a consequence of hydrolytic decomposition.

Another morphological feature of this horizon is the development of a characteristic greyish film or coating on the surface of the structural elements, which often tend to be prismatic or coarsely polyhedral. This appearance, known as "gleying," is very characteristic, and the horizon is known as the gley or glei horizon. Whilst the rusty mottling generally accompanies gleying, it is not invariably present, and incipient gleying can be seen, particularly in sandy soils without observable rust mottling.

In the permanently anaerobic horizon below the lowest level of the water-table, there is no root penetration and the organic matter content is very low. Owing to the prevalence of reducing conditions, the material has a bluish-grey, bluish-black, or greenish-grey colour due to the presence of ferrous compounds, among which ferrous sulphide, FeS, and vivianite, Fe₇(PO₄)₃, may be present.

In soils where wetness is due to the impervious character of the profile material, the three zones are less clearly defined. In many cases, as for example in certain heavy clay soils, the wetness is confined to the top few feet which overlie a compact dry mass of clay into which moisture does not penetrate. In such soils, the profile may become almost completely dried out in the summer and the zone of permanently reducing conditions is absent. The zone of gleying and rust mottling may also be more diffuse.

Very great variations may occur in the character and thickness of the humus layer and in the development of
mottling and gleying in the gley horizon through the deposition of materials from the ground water. The general grey-
ness of the soil horizons, the development of prismatic or polyhedral structure, the occurrence of gleying, and the
presence of secondary deposits are all indications of this soil group.

The natural vegetation of meadow soils is generally of a gramineous, juncaceous, or caricaceous type. They occur
also under wet woodland vegetation, as in the so-called taiga of Siberia. When the ground-water level approaches
more nearly to the surface and permanently wet conditions prevail, the tendency is towards peat formation with
the development of a gley podsol or peat podsol profile, consisting typically of up to 1 foot of peat underlain by grey
mottled subsoil. This type passes into the peat soils.

Gley soils are common in depressions and hollows in regions of podsolic soils. In passing from an upland to the
bottom of a depression in such regions, a regular succession of profiles can frequently be observed. In the upland, with
free drainage and scanty accumulation of peaty organic matter, iron podsols may occur. In the moister lower levels,
the humic layer becomes thicker, and gley podsols and peat podsols result. In still lower ground, the water-table is
sufficiently near the surface to give the gley horizon characteristic of a meadow profile, which gives place in turn to a
peat profile. Gley podsols may result through impedance consequent on the development of ortstein in podsol profiles.

Gley soils are essentially local in character and depend, in the main, on topographical conditions which influence
the movement of underground water. Most typically they are developed in hollows, but where drainage is im-
peded by impervious strata on slopes, the typical characters of the gley profile may also be developed.

Gley soils occur throughout middle, northern, and western Europe, and, indeed, throughout all the humid
regions. In Europe, they attain their greatest development in the Priepet marsh region of Poland.
Analogous conditions in more arid regions may give rise to saline and alkaline soils. There are also characteristic soils associated with high ground-water in the tropics. These will be considered at a later point.

Finland. B. Frosterus\(^\text{a}\) gives the following description of a gley soil in Finland:

A\(_1\).  0-23 cm. Brownish-grey sandy peat.
A\(_3\).  27-32 cm. Yellowish-brown plastic clay passing sharply into
B.  32-45 cm. Ochreous plastic clay passing gradually into
B-G. 45-120 cm. Grey clay with abundant rusty threads and concretions arranged in three horizontal layers at 45, 50, and 80 cm., respectively.
G\(_1\). (Gley horizon). 120-175 cm. Grey clay with scattered rusty threads and prismatic structure.
G\(_2\). 175-195 cm. Prismatic clay with superficial rusty coating on the individual columns into which the horizon is dissociated by vertical cracks.
CG. Blue clay, occasionally inky black containing vivianite.

Germany. The soils known in parts of Germany as "Molkenböden" may be grouped with the meadow soils. K. V. Von Falckenstein\(^\text{b}\) describes soils of this kind developed from the middle Bunter sandstone under conditions of impeded drainage in the Weser valley. The following profile is typical:

0-3 cm. Litter layer.
3-16 cm. Bleached humous layer.
16-50 cm. Compact layer with yellow streaks becoming more stony below.

Analyses of the successive layers show marked differences from podsols, for there is no leaching of alkalies and alumina and only a slight downward movement of ferric oxide. The bleaching in the surface horizon is considered to be the effect of reduction of ferric to ferrous compounds.

Scotland. W. G. Ogg\(^\text{c}\) describes a gley soil developed on Silurian shale boulder clay in the Moorfoot Hills. The altitude is 1,250 feet, the mean annual rainfall 35-40 in., and the vegetation *Juncus communis* and *Carex Goodenovii*
dominant, with *Holcus lanatus*, *Trifolium repens*, and *Anthoxanthum odoratum* abundant.

(1) 0-5 cm. Mat of roots and moss.
(2) 5-15 cm. Brownish-grey silty loam; wet; no definite structure; few stones; roots common; some brown patches along root-channels.
(3) 15-22 cm. (Variable.) Grey silty loam with brown patches and streaks; wet; few stones; roots common; brown staining along root-channels.
(4) 22-35 cm. Brown silty loam with greyish tinge; wet; slightly gritty; few stones; much brown mottling and staining along root-channels.
(5) 35-55 cm. Greyish-brown silty loam with sandy patches; wet; stones common; blackish particles or concretions; roots penetrate; brown staining along root-channels.
(6) 55 cm.+ Stony silty loam; grey-brown with brown patches; very wet; gritty.

The principal analytical data are given in Table XXXII.

**Table XXXII.—Analytical data for Scottish Gley soil**

<table>
<thead>
<tr>
<th>Depth (in cm.)</th>
<th>0-5</th>
<th>5-15</th>
<th>15-22</th>
<th>22-35</th>
<th>35-55</th>
<th>55 +</th>
</tr>
</thead>
<tbody>
<tr>
<td>pH</td>
<td>5.2</td>
<td>5.3</td>
<td>5.6</td>
<td>5.4</td>
<td>5.8</td>
<td>5.5</td>
</tr>
<tr>
<td>Ignition loss</td>
<td>18.4</td>
<td>12.6</td>
<td>7.3</td>
<td>5.2</td>
<td>4.1</td>
<td>3.6</td>
</tr>
<tr>
<td>Clay SiO₂/Al₂O₃</td>
<td>2.96</td>
<td>2.85</td>
<td>2.98</td>
<td>2.89</td>
<td>2.79</td>
<td>2.29</td>
</tr>
<tr>
<td>,, SiO₂/Fe₂O₃</td>
<td>2.44</td>
<td>2.32</td>
<td>2.37</td>
<td>2.28</td>
<td>2.12</td>
<td>1.82</td>
</tr>
<tr>
<td>,, Al₂O₃/Fe₂O₃</td>
<td>4.69</td>
<td>4.38</td>
<td>3.89</td>
<td>3.74</td>
<td>2.41</td>
<td>3.87</td>
</tr>
</tbody>
</table>

**Wales.** Soils developed under conditions of impeded drainage are commonly encountered throughout Wales, but occur more frequently in districts of high rainfall.

We may distinguish two types, namely, soils in which the impedance of drainage is due to the high level of the water table, and soils in which impedance is due to the occurrence of a relatively impervious stratum in the subsoil.

As an example of the first type, a profile from near Bangor, N. Wales, may be described. The parent material
is alluvium derived from Ordovician Shale. The soil is described for soil survey purposes as the Conway series.

The profile is as follows:

<table>
<thead>
<tr>
<th>Depth</th>
<th>Description</th>
<th>Clay %</th>
<th>Organic Carbon %</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>o-8 in.</td>
<td>Grey heavy loam</td>
<td>28.9</td>
<td>5.30</td>
<td>6.1</td>
</tr>
<tr>
<td>8-20 in.</td>
<td>Grey mottled-brown clay</td>
<td>35.7</td>
<td>2.47</td>
<td>5.8</td>
</tr>
<tr>
<td>&gt;20 in.</td>
<td>Grey mottled-brown heavy clay</td>
<td>45.9</td>
<td>1.06</td>
<td>5.7</td>
</tr>
</tbody>
</table>

The composition of the clay fraction in the successive horizons is as follows:

<table>
<thead>
<tr>
<th>Horizon</th>
<th>SiO_2/Al_2O_3</th>
<th>SiO_2/Fe_2O_3</th>
<th>Al_2O_3/Fe_2O_3</th>
</tr>
</thead>
<tbody>
<tr>
<td>o-8 in.</td>
<td>2.70</td>
<td>2.40</td>
<td>8.00</td>
</tr>
<tr>
<td>8-20 in.</td>
<td>2.73</td>
<td>2.43</td>
<td>8.01</td>
</tr>
<tr>
<td>&gt;20 in.</td>
<td>2.62</td>
<td>2.26</td>
<td>6.28</td>
</tr>
</tbody>
</table>

This soil is typical of considerable areas of wet lowland soils in Wales. The vegetation is generally grass and rushes, together with miscellaneous water-loving plants in the wetter situations. All stages of transition to lowland peat soils may be observed.

The data for the composition of the clay fraction are instructive. In spite of the occurrence of much brown mottling in the lower horizons, there does not appear to be any considerable eluviation of ferric oxide. The low figures for this constituent throughout the profile are reminiscent of the A horizon material of a podsolic profile and it may be that the whole profile consists of A horizon material removed by former erosion and deposited in its present low-lying situation.

The iron content of the clay of bottom soils is not always so low as in the profile described, and the grey colour may be partly attributed to the existence of iron in the ferrous state consequent on the reducing conditions which commonly obtain in such soils.

An example of a soil with drainage impeded by an impervious stratum is furnished by a profile from Llanfairpwl, Anglesey, mapped under the name of the Gesail series for survey purposes. It occurs under grass vegetation. The
parent material is micaceous schist boulder clay and the profile is as follows:

<table>
<thead>
<tr>
<th>Depth</th>
<th>Description</th>
<th>Clay %</th>
<th>Organic Carbon %</th>
<th>pH</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-10 in.</td>
<td>Greyish loam</td>
<td>20.7</td>
<td>3.6</td>
<td>6.9</td>
</tr>
<tr>
<td>10-20 in.</td>
<td>Greyish mottled micaceous loam</td>
<td>15.7</td>
<td>0.8</td>
<td>8.0</td>
</tr>
<tr>
<td>20-30 in.</td>
<td>Grey and brown mottled micaceous heavy loam</td>
<td>14.3</td>
<td>1.4</td>
<td>8.0</td>
</tr>
</tbody>
</table>

A certain amount of manganese dioxide occurs in the lower horizons, and the pH values determined by the quinquhydrone method are rather too high.

The composition of the clay fraction in the successive horizons is as follows:

<table>
<thead>
<tr>
<th>Horizon</th>
<th>SiO₂/Al₂O₃</th>
<th>SiO₂/R₂O₅</th>
<th>Al₂O₃/Fe₂O₅</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-10 in.</td>
<td>2.50</td>
<td>2.05</td>
<td>4.53</td>
</tr>
<tr>
<td>10-20 in.</td>
<td>2.63</td>
<td>1.99</td>
<td>3.11</td>
</tr>
<tr>
<td>20-30 in.</td>
<td>2.66</td>
<td>2.02</td>
<td>3.15</td>
</tr>
</tbody>
</table>

There is a certain eluviation of ferric oxide, but there is no very decided difference between the composition of the clay fraction in the different horizons.

In the two profiles described above, the mottled sub-surface horizons may be equated with the gley horizons of continental writers. It should be noted that the brown or reddish-brown appearance is not due to a deposition throughout the mass of the soil as in the B horizon of podsolic profiles, but rather to superficial deposition on the structural elements. This is reflected in the comparatively slight enrichment in ferric oxide shown by the analytical figures for the clay fraction.

One feature of meadow soil profiles is the fairly sharp decrease in organic matter content in passing from the turf layer. The moisture conditions are such as to restrict root development to the surface, so that comparatively little organic matter is present in the layers subjected to waterlogging.

England. Certain of the soils of central and eastern England derived from heavy clay formations might be
appropriately grouped with the meadow soils. The following profile at Lolworth, Cambridgeshire, on Kimmeridge Clay is described by F. Hanley:

0-9 in. Dark brown clay loam with flints: mottlings.
9-21 in. Grey-brown clay with light brown mottlings.
21-40 in. Lead-grey clay with brown mottlings.
>40 in. Lead-grey clay; yellow seams in upper portion and pockets of crystalline gypsum.

The analytical figures show an almost complete leaching of carbonates and a marked accumulation of gypsum below 40 inches. The soil contains 32.8% clay, but the lower horizons have from 60-70% clay.

United States. Gley soils are of common occurrence throughout the Union and a perusal of the publications of the Soil Survey will afford many instances. A typical example is the Elkton series found in many parts of Maryland and the adjoining states. These soils occur in flat areas or small depressions and near the heads of streams and are derived from unconsolidated clay sediments. A description by S. O. Perkins and H. B. Winant of a profile near Stevensville, Md., is as follows:

A. 0-5 in. Light grey silt loam, compact when dry but friable under favourable moisture conditions. Floury feel.
B₁. 5-6 in. Light grey heavy silt loam, faintly mottled with rust-brown.
B₂. 8-40 in. Light grey silt clay, mottled rust-brown. Slightly plastic with thin layers of grey fine sand and lenses of heavy plastic clay.
C. 40-50 in. Light grey silt loam or heavy fine sandy loam, mottled brownish-yellow. Very variable.

GLEY PODSOLS AND PEAT PODSOLS

When drainage impedance is superimposed on podsol development a gley podsol results. This is accompanied by an increase in the peaty layer which eventually leads to the development of a peat podsol profile. The drainage impedance may be due to ortstein development or to the formation of a clay pan by mechanical illuviation. Another possibility is swamping due to deforestation. I. V. Tiurin,
V. A. Burenkov, and A. S. Masslov give an example of this from Russia. Where re-planting takes place the swamping gradually disappears in 15-20 years as the new stand of trees grows up. In the absence of re-forestation, it may be presumed that the original podsol character of the profile would become modified.

**Scotland.** A. Muir has described numerous profiles of gley podsolic and peaty gley podsolic (peat podsol) soils. The following example of a peaty gley podsol may be quoted. It occurs in lower Strathspey in the Teindland Forest. The parent material is boulder clay (till) and the vegetation *Calluna, Scirpus, Erica tetralix, Carex spp., Cladonia spp., Sphagnum compactum* in hollows, and stunted pine scrub. The altitude is 700 feet.

A<sub>0</sub>. 0-7 (10) cm. Surface root mat over black well-decomposed peat. Stones occasional.

A<sub>2</sub>-G. 7 (10-17) cm. Dark grey, sandy; humus staining throughout. Stony and free; many roots.

B<sub>1</sub>. 17-23 cm. Brownish-black humus accumulation layer, sandy and stony, roots numerous. Underlain by discontinuous hardpan which is quite thin.

B<sub>2</sub>. 23-38 cm. Iron-stained sandy, stony, roots occasional, structureless, slightly compacted yet appears free. Slight humus staining.

B<sub>3</sub>. 38-53 cm. Lighter in colour than previous layer, no humus staining visible. Fairly compact, stony and structureless.

C. 53-80 cm. Purplish-red till, stony, structureless; slightly compacted.

The principal analytical data are:

<table>
<thead>
<tr>
<th></th>
<th>A&lt;sub&gt;2&lt;/sub&gt;-G</th>
<th>B&lt;sub&gt;1&lt;/sub&gt;</th>
<th>B&lt;sub&gt;2&lt;/sub&gt;</th>
<th>B&lt;sub&gt;3&lt;/sub&gt;</th>
<th>C</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ignition loss %</td>
<td>4.70</td>
<td>13.53</td>
<td>9.36</td>
<td>2.80</td>
<td>2.59</td>
</tr>
<tr>
<td>Clay SiO&lt;sub&gt;2&lt;/sub&gt;/Al&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;3&lt;/sub&gt;</td>
<td>2.25</td>
<td>1.75</td>
<td>1.80</td>
<td>1.48</td>
<td>1.98</td>
</tr>
<tr>
<td></td>
<td>SiO&lt;sub&gt;2&lt;/sub&gt;/R&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;3&lt;/sub&gt;</td>
<td>1.89</td>
<td>1.15</td>
<td>1.25</td>
<td>1.25</td>
</tr>
<tr>
<td></td>
<td>Al&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;3&lt;/sub&gt;/Fe&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;3&lt;/sub&gt;</td>
<td>5.25</td>
<td>1.92</td>
<td>2.27</td>
<td>5.44</td>
</tr>
</tbody>
</table>

**Wales.** Gley podsols and peat podsols are widely distributed throughout Wales. In many cases they occur in situations where heath vegetation on freely drained soils has been succeeded by moorland vegetation with drainage im-
I. Pakihi (Gley-podsol) profile, Nelson Province, New Zealand.

II. Soloti profile in lacustrine clay; N. Dakota, U.S.A.

PLATE VII.
Hydromorphic Soils and Peats

Pedance and peat formation. Such soils are common at high altitudes. The following profile is described by T. Waśowicz.

N.E. of summit of Foel Fras at 3,000 feet, under vegetation of Juncus squarrosus, Calluna vulgaris, Molinia coerulea, mosses, Empetrum nigrum, Vaccinium myrtillus and Nardus stricta.

A. 0-90 cm. Dark brown wet fibrous peat.
A. 90-100 cm. Grey silty loam with bleached gravel.
B. 100-105 cm. Rusty-brown gravelly loam, the intensity of the colour increasing with depth, passing to shattered rock.

The analytical data are as follows:

| Exch. CaO Clay Org. Carbon Clay fraction |
|---|---|---|---|---|---|
|   | % | % | SiO₂/Al₂O₃ | SiO₂/R₂O₃ | Al₂O₃/Fe₂O₃ |
| A | 3.88 tr. 6.6 3.0 2.41 2.13 7.69 |
| B | 4.47 tr. 9.0 2.2 1.19 0.75 1.68 |

New Zealand. T. Rigg describes certain soils, termed Pakihi soils, occurring in the Nelson province which appear to be essentially gleysols.

The following is a description of a Pakihi soil (see Plate V) under Gleichenia dicarpa, Cladium teretifolium, Juncus planifolius, LycoPodium and Drosera spp.

1. 4 in. of raw humus.
2. 3½ in. of grey sandy silt.
3. 5 in. of brownish fine sand merging into coarse sand.
4. 1 in. of coarse grit.
5. 4 in. of pan cemented with iron and humus compounds over gravel and stones stained in neighborhood of pan.

Vlei* Soils

This name has been given to certain African soils which occur in depressed or basin-shaped areas subject to seasonal wetness. They may be black, brown, or grey in colour and often show mottling due to the prevalence at certain seasons.

*This Afrikaans word, pronounced *fly, may be conveniently used for the tropical and sub-tropical analogues of the gleys soils. The alternative term *mbuga, used by Milne, might serve but is less euphonious.
of waterlogged conditions. Ironstone concretions (murrum) are frequently present, and calcium carbonate nodules are also encountered.

Kenya. D. S. Gracie\textsuperscript{16} gives the following description of a profile in the South Thomson’s Falls district of Kenya:—

0–9 in. White soil with mottling. Poor accumulation of organic matter at the surface. A band of iron concretions separating this layer from the next.

9–31 in. Compacted organic clay layer, whitish at first, but shading into black and becoming brown at its lower limit.

31–60 in. Yellowish material, soft and friable in the first six inches, then becoming harder. Large calcium carbonate concretions present.

The surface soil has a pH of 4.11 and contains 0.183% nitrogen.

Tanganyika. G. Milne\textsuperscript{17} describes a typical vlei soil in a saucer-shaped depression at Muhesa, Tanganyika. The altitude is 600 feet and the rainfall 50 inches, with maxima in March-May and November-December. The original vegetation was tall dense grass. The profile is:—

0–4 in. Friable grey-black surface soil.

4–10 in. Inky black sticky subsoil. Below 10 in. Mottled yellow soapy clay to 11 feet.

There are many ferruginous concretions through the subsoil and traces of calcium carbonate from 18 inches downwards. The following data for the Muhesa vlei soil were obtained in the writer’s laboratory (Table XXXIII).

\begin{table}
\small
\centering
\begin{tabular}{|l|c|c|c|c|c|c|c|}
\hline
Depth & Clay & Organic & pH & Clay Fraction & SiO$_2$/Al$_2$O$_3$ & SiO$_2$/R$_2$O$_3$ & Al$_2$O$_3$/Fe$_2$O$_3$
\hline
0-4 in. & 27.6 & 2.99 & 7.6 & 2.99 & 2.47 & 1.62
4-8 in. & 31.9 & 2.56 & 6.9 & 2.92 & 2.36 & 1.66
26-32 in. & 45.5 & 0.85 & 6.3 & 2.15 & 1.79 & 5.00
\hline
\end{tabular}
\end{table}
Cuba. An example from Cuba of what would appear to be an analogue of the vlei soils is furnished by the Mabo series of H. H. Bennett and R. V. Allison. These soils are developed on outwash from siliceous rocks in regions of flat topography with poor drainage. The profile consists of:

- 0-3 in. Grey silt loam or fine sandy loam with some rusty brown mottling. Acid reaction.
- 3-8 in. Friable yellowish-grey or mottled pale yellow and grey material.
- 8-12 in. Pale yellow and friable material with abundance of perdigón (concretions).
- >12 in. At less than two feet. Mottled red and bluish-grey plastic impervious heavy clay.

The Coxville series of Cuba is similar to the Mabo series, but concretions are comparatively rare.

S. Rhodesia. H. B. Maufe describes a vlei soil derived from dolerite at Salisbury, S. Rhodesia. The profile consists of:

- 0-9 in. Black soil.
- 28-43 in. Yellowish-brown clay with rounded blocks of fresh dolerite.

The top soil is coarsely granular, but the subsoil has a columnar or prismatic structure. Although the pit in which this profile was seen was waterless at the end of the dry season, the bottom part was distinctly wet, indicating poor drainage conditions. In a better drained situation in the immediate locality, the same rock gives rise to a red soil.

LATERITIC SOILS

These soils, developed under conditions of intermittent waterlogging either in contemporary or past times, have some claim to be included in the present chapter. In view of their widespread occurrence in the tropics, however, they have been retained in Chapter XIV.

GROUND-WATER PODSOILS

This term is given to a type of profile, developed generally in a light-textured material with a high water-table,
in which there is a strong bleaching of the A horizon and an equally marked deposition in the B horizon near ground-water level. The Leon series of Florida is described by C. E. Kellogg as having "an exceedingly thin A horizon underlain by an acid, leached A₂ horizon of light grey fine sand, slightly stained with humus in the upper part. The B horizon is sandy, but darkly stained and cemented, largely with organic material, into a hard ortstein. The roots of plants are confined to the A horizon. . . . The water-table stood at about 3 feet where the B horizon merges into wet loose sand." Certain soils in San Domingo, described by E. Balzarotti as podsolized, may be similar and owe their characters to ground-water effects. They occur under savannah.

**ALLUVIAL AND SWAMP SOILS**

Soils derived from marine, estuarine, and river alluvial deposits can show all stages of development according to their age. Where time has sufficed, development can occur on alluvial material just as on any other kind of parent material. It would therefore be incorrect to describe a podsol developed on alluvium as an alluvial soil. The term is best restricted to juvenile soils with undeveloped profiles in which the soil characters are dominated by the parent material.

Agriculturally, these soils are of considerable importance. They are found in most countries and are generally of high fertility. Examples are the polder soils of Holland and the delta soils of Egypt. In many cases these have been reclaimed from tidal water by means of dykes. A considerable proportion of Holland has been thus won from the sea. The soils of older reclamation have undergone considerable changes. In the early years there is a leaching out of soluble salts, followed by a gradual loss of calcium carbonate and fall in base status. With skilful management, including the regulation of water conditions by drainage, these soils are very productive. In England the fertile agricultural soils of S. Lincolnshire are of this class.
In many parts of the tropics, mangrove swamps occur in coastal districts. These are sometimes reclaimed and used for paddy (rice) cultivation. Tropical swamp soils have, as yet, been insufficiently studied; but they present many points of interest. H. C. Doyne and R. R. Glanville\(^2\), in a study of mangrove swamp soils in Sierra Leone, found pH values of less than 3 in recently cleared soils. This intense acidity appears partly to be due to the presence of free hydrochloric acid from the hydrolysis of soluble ferrous salts. (See also p. 159.) The acidity of the top soil decreases after a few years of cultivation, but remains in the subsoil.

**BLACK PRAIRIE SOILS**

There is a group of soils in Canada and the United States, termed *black prairie soils*, which appear to be essentially meadow soils. In a description of Saskatchewan soil profiles, A. H. Joel\(^3\) mentions a type of "black park land" profile, consisting of a "black or almost black A horizon of mixed loose and fine granular structure; a B horizon of some shade of brown, heavier and more compact than the surface, frequently somewhat columnar in structure and usually fine to medium granular when crumbled; a \(B_2\) horizon of grey to light grey colour, frequently with a yellowish or brownish tinge and flowing to granular structure (zone of carbonate accumulation); and a C horizon of grey to dark grey colour mottled with flecks or stains of reddish-brown ferric oxide and frequently with concretions or splotches of calcium carbonate, calcium sulphate or both. The \(C_2\) horizon is usually a more uniform dark grey or somewhat bluish dark-grey." A shallow and a deep phase are distinguished, the deep phase occupying lower elevations. Whilst the shallower phase is considered to be well drained and may be a type of chernozem, the deep phase shows evidences of poor drainage conditions in iron stains and mottlings. The black layer varies from about 3 to 8 inches in the deep phases. It is thus rather shallower
than in typical chernozems to which, on the grounds of carbonate accumulation, these soils might be assigned.

From this circumstance and the analytical evidence of mechanical and chemical eluviation, Joel excludes them from the chernozem group. It would appear that the deeper phases, at least, must be assigned to the meadow soil group.

Black soils in Alberta, described by F. A. Wyatt and J. D. Newton\textsuperscript{24}, show a somewhat similar profile, with an organic layer of about 12 inches in thickness overlying about 30 inches of grey columnar to granular clay loam or clay, with carbonate accumulation in the lower 6 inches. This is underlain by light yellow clay with occasional iron stains. The data for the nitrogen content of successive layers, both of the Saskatchewan and the Alberta black soil profiles, show a sharp decrease in organic matter in passing from the dark layer. This is markedly different from the gradual decrease in organic matter with depth observed in the chernozems and suggests the effect of impeded drainage conditions.

Considerable discussion has centred round the so-called prairie soils of the Middle-West of the United States. Whilst many of the European workers assign them to the meadow soils, C. F. Marbut\textsuperscript{25}, whilst admitting that, in the earliest stage of topographical development, impeded drainage may have obtained, considers the present drainage to be unimpeded. There are, however, areas in South-Central Iowa, North-Eastern Missouri, and parts of Southern Illinois where meadow soils occur owing to the survival of drainage impedance.

China. C. F. Shaw\textsuperscript{26} refers to a class, known as Sajong soils, occurring in the central plains of China. They occur with an almost level topography. They show horizons of calcium carbonate accumulation at depths varying from one to six feet. From the description of the drainage conditions, it seems likely that they belong to the meadow soils rather than to the chernozem group. The calcium carbonate
would thus represent a deposit in the vicinity of a water-table.

S. AFRICAN BLACK TURF SOILS

There is a class of soils occurring typically in the Transvaal, but also found with no essential difference in Southern Rhodesia, to which the name black turf has been applied. The profile consists of a variable depth, commonly about three feet, of grey-black or brownish-black clay, granular when dry but waxy when wet. Calcereous concretions occur in the lower part of the layer, increasing in number to form eventually a friable calcereous layer, below which there is a transition to decomposing norite rock. In some cases the calcereous layer is absent and the black layer rests directly on the decomposing rock.

B. de C. Marchand\textsuperscript{27}, in a discussion of the origin of these soils, controverts the theory advanced by C. F. Marbut\textsuperscript{28}, that the black turf soils belong to the chernozem group, as might be expected from their black colour and the presence of a layer of carbonate accumulation. He shows that their occurrence is strictly associated with the outcrop of norite rock, whilst, under identical climatic and topographical conditions, diabase yields a completely leached friable red loam. They cannot be described as vlei soils since they are not limited to areas of depressed topography. The critical factors in their formation appear to be impeded percolation, consequent on the impervious character of the clay formed by the weathering of the norite, and their high base status, which results in the development, as elsewhere in hot climates, of a dark type of organic matter. The dark colour of these soils may convey an erroneous impression of their organic matter content, which is not generally higher, and may in some cases be actually lower, than that of the adjoining red loams.

The clay fraction of the black turf soils reflects the restriction of percolation, for it is markedly siliceous in character. Whilst the clay fraction of the leached friable red
loams has a silica-sesquioxide ratio of 2.0 or less, the clay fraction of the black turfs has a silica-sesquioxide ratio of 3.0 or more. The high base-status is also contributory.

C. R. Van de Merwe\(^{29}\) considers these soils to be morphologically akin to the Indian regur and the black cotton soils of Kenya.

CONDITIONS OF DEVELOPMENT OF GROUND-WATER SOILS

In the present chapter, we have discussed soils with impeded drainage from regions with widely differing climatic conditions. Whilst they occur more frequently in humid regions such as North-West Europe, they may be found also under comparatively dry climates such as that of the Canadian prairies. The common characters that distinguish them are greyish to black surface horizons sharply defined from the sub-surface horizons, which, by their gleying, iron stains, mottlings, or concretions, and also, in many cases, by the presence of gypseous or calcareous horizons, give evidence of the alternation of aerobic and anaerobic conditions due to variable ground-water levels. The deposition of hydrated ferric oxide may often reach notable proportions, as in the soft bog iron-ore of northern lands and the harder "murram" or "mocarrero" of the tropics.

Soils developed under conditions of impeded drainage may, in some cases, show resemblance to the black earths. The relatively high content of organic matter often developed under neutral or alkaline conditions may result in a dark colour in the surface horizons. Further, the drainage impedance and the high base-status both tend to the development of a siliceous type of weathering complex analogous to that of the chernozem group. Finally, the presence of a ground-water table may in many cases be associated with the deposition of calcium carbonate in the adjacent horizons.

Meadow and vlei soils are, however, distinguished from the chernozems by the absence of the granular structure and
by the occurrence of rusty mottlings or streaks, indicating the alternation of oxidative and reductive conditions.

In the soils of this group examined by the writer, the mineral colour of the surface horizons is generally grey and never red or reddish-brown. Such analyses as have been made show the clay complex to be of a siliceous type with silica-sesquioxide ratios in excess of 2.0. It would be of interest to determine whether this is a constant feature.

PEAT SOILS

Peat soils are distinguished by the high proportions of organic matter which they contain. They are soils in which the residues of natural vegetation are humified rather than oxidized. A distinction is sometimes drawn between peat soils and mineral soils, but it is somewhat difficult to assign a limiting content of organic matter above which a soil is to be considered a peat. Whilst 15% of organic matter may suffice to give a pronounced peaty character to a light sand, a heavier soil may contain up to 20% of organic matter and still retain the appearance and character of a mineral soil. Typical peats, however, generally contain at least 50%, and may in some cases consist almost entirely of organic matter.

The humification that results in the formation of peat is, except in the case of heath peat and forest peat, of an anaerobic character, and takes place either under water, as in the formation of fen peat, or in situations where waterlogging and consequent exclusion of air is the rule. From this it follows that peat is most readily developed under cool humid climates. But whilst they are commonly found under such conditions, peats can occur in the tropics, as for example, in mangrove swamps. Notable deposits of peat occur under sub-tropical conditions in the Everglades of Florida.

Typically, peats are formations of temperate and cool climates. They are more widespread in northern regions, partly because of the extensive spreads of glacial drift with its associated lakes, and partly because the scanty evapora-
tion associated with low temperatures results in a greater proportion of the rainfall being available as soil-water.

Most of the peat of Northern Europe and America has been formed in basins formerly occupied by lakes. Existing deposits represent different stages in a process which has proceeded since the close of the Glacial Period and is still in progress. The relationships of the different types of peat to each other will be evident from a description of this process.

The first stage of basin peat formation takes place under lacustrine conditions. In the shallow water near the shore, associations of such plants as _Phragmites_, _Carices_, _Juncus spp._, and _Equisetaceae_ become established. The residues of these plants accumulate at the bottom, where they are mixed to some extent with lake mud and undergo anaerobic decomposition. The littoral association encroaches more and more on the free water surface until the lake is completely filled up by the humified residues of the plants which have inhabited it, mixed with a certain proportion of mineral matter carried by the stream feeding the lake and deposited as sediment. Examples of all stages of this process may be seen in any region of shallow lakes. The peat thus formed is known as low-moor or fen peat ("Niederungs-moor" or "Flachmoor").

With the disappearance of the free water surface and the accumulation of humified material, a gradual change in the type of vegetation sets in: mosses (e.g., _Hypnum_), _Calluna vulgaris_, _Erica tetralix_, _Molinia caerulea_, and finally trees, such as alders (_Alnus_), birches (_Betula nana_ and _pubescens_) and pines (_P. sylvestris_ and _uliginosus_) make their appearance, whilst the earlier lacustrine plants die out. The peat accumulated from the residues of the new vegetation is now of a different type and may be recognized, when this stage is past, by the presence of tree remains.

The conditions at the close of the fen stage and before the establishment of tree growth are relatively dry, owing to the growth of the peat deposit above the original water level. The establishment of tree growth, however, by protecting
the surface layers from losses by evaporation promotes uniformly moist conditions allowing the growth of sphagnum moss. Meanwhile, there has been a progressive decline in plant nutrient status. The plants which contribute to fen peat are relatively rich in plant nutrients and grow in water containing considerable proportions of mineral matter in solution and suspension. As the peat grows above the fen level and contributions from telluric waters are diminished, the type of vegetation which prevails consists of plants which make scanty demands for nutrients from the medium on which they grow. The peat formed from their decay becomes progressively poorer in nitrogen and mineral constituents. This, and the wetness consequent on the high retentivity of the sphagnum for moisture, eventually results in the annihilation of tree growth, and the peat deposit now enters on its final phase as a high-moor peat or bog (Hochmoor) in which the principal plants are sphagnum species and cotton grass (*Eriophorum*).

In a fully developed profile, a section will disclose a complete succession from the beginning of the fen peat stage. W. Bersch\(^{30}\) gives the following typical example:—

1. Vegetation layer.
2. Younger moss peat, 140 cm.
3. Light coloured transition layer (heath), 20 cm.
4. Older moss peat, 170 cm.
5. Woody peat, 85 cm.
6. Fen peat (*Phragmites* and *Carices*), 70 cm.
7. Peat mud, 15 cm.

Horizons 1 to 4 represent the high-level peat; horizon 5, the transition (woodland) stage; and horizons 6 and 7, the fen peat. The transition layer between the older and the newer peat moss is generally observed in continental high level peats and is considered to represent a period of relatively dry conditions involving an intermission of sphagnum peat formation, and the establishment of heath.

Peat deposits may reach to a depth of 50 feet or even more. The rate of accumulation has been variously esti-
mated. Whilst the vertical rise of the surface may amount to 1-2 cm. per annum, the actual growth of solid peat must be considerably less. An estimate may be gathered from an example at Laibach, given by Bersch, where a Roman road of about the first century A.D. was found below 120 cm. of peat. Assuming 1,800 years as the upper limit of time, the rate of accumulation is about 0.7 mm. per annum.

Lacustrine or fen peat is markedly different in character and composition from sphagnum or bog peat, for not only are the plants from whose remains it has been formed richer in ash constituents, but there is also an admixture of sedimentary material of terrestrial origin.

S. A. Waksman and K. R. Stevens examined samples of lacustrine and sphagnum peats and their results given in Tables XXXIV and XXXV show clearly the contrast in composition between the two types. (Cf. pp. 203 et seq.)

**Table XXXIV. — Percentage Composition of Several Horizons of a Low-moor Peat Profile from Newton, New Jersey, U.S.A., on Basis of Dry Matter (Waksman and Stevens)**

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Ether-Soluble Fraction</th>
<th>Water-Soluble Fraction</th>
<th>Hemi-Celluloses</th>
<th>Cellulose</th>
<th>Lignin</th>
<th>Crude Protein</th>
<th>Ash</th>
</tr>
</thead>
<tbody>
<tr>
<td>$A_1$</td>
<td>0.66</td>
<td>3.08</td>
<td>10.31</td>
<td>0</td>
<td>38.35</td>
<td>22.48</td>
<td>13.22</td>
</tr>
<tr>
<td>$A_2$</td>
<td>1.10</td>
<td>1.24</td>
<td>8.95</td>
<td>0</td>
<td>50.33</td>
<td>18.72</td>
<td>10.13</td>
</tr>
<tr>
<td>$A_3$</td>
<td>0.49</td>
<td>2.31</td>
<td>7.02</td>
<td>0</td>
<td>57.83</td>
<td>14.81</td>
<td>10.15</td>
</tr>
<tr>
<td>$A_4$</td>
<td>0.78</td>
<td>1.14</td>
<td>7.51</td>
<td>0</td>
<td>42.10</td>
<td>19.81</td>
<td>15.00</td>
</tr>
<tr>
<td>Lake peat</td>
<td>0.67</td>
<td>0.81</td>
<td>12.14</td>
<td>0</td>
<td>33.25</td>
<td>19.38</td>
<td>24.87</td>
</tr>
<tr>
<td>Bottom mud</td>
<td>0.36</td>
<td>1.24</td>
<td>5.92</td>
<td>0</td>
<td>15.62</td>
<td>9.81</td>
<td>59.55</td>
</tr>
</tbody>
</table>

The first point to be noticed is the contrast in the ash contents. Whilst, in the low-moor peat, the ash varies from about 10 to 25%, in the high-moor, the ash is between 1 and 2%, rising in the woody peat, which represents a transition to low-moor, to over 5%.

Secondly, whilst cellulose is completely decomposed under the conditions obtaining in low-moor, the high-moor
Table XXXV.—Percentage composition of several horizons of a high-moor (Sphagnum) peat profile from Oldenburg, Germany, on basis of dry matter (Waksman and Stevens)

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Ether-Soluble Fraction</th>
<th>Alcohol-Soluble Fraction</th>
<th>Hemim Celluloses</th>
<th>Cellulose</th>
<th>Lignin</th>
<th>Crude Protein</th>
<th>Ash</th>
</tr>
</thead>
<tbody>
<tr>
<td>Younger sphagnum 1</td>
<td>3.08</td>
<td>...</td>
<td>16.88</td>
<td>19.44</td>
<td>34.04</td>
<td>5.23</td>
<td>1.72</td>
</tr>
<tr>
<td>,, ,, 2</td>
<td>6.12</td>
<td>...</td>
<td>11.09</td>
<td>13.62</td>
<td>49.56</td>
<td>5.11</td>
<td>1.85</td>
</tr>
<tr>
<td>,, ,, 3</td>
<td>3.96</td>
<td>...</td>
<td>16.24</td>
<td>19.91</td>
<td>38.26</td>
<td>6.58</td>
<td>1.50</td>
</tr>
<tr>
<td>,, ,, 4</td>
<td>5.02</td>
<td>...</td>
<td>15.44</td>
<td>16.38</td>
<td>40.72</td>
<td>6.11</td>
<td>2.36</td>
</tr>
<tr>
<td>,, ,, 5</td>
<td>4.99</td>
<td>...</td>
<td>15.12</td>
<td>17.14</td>
<td>43.82</td>
<td>5.10</td>
<td>1.82</td>
</tr>
<tr>
<td>Heath peat ...</td>
<td>7.60</td>
<td>...</td>
<td>8.44</td>
<td>9.76</td>
<td>54.38</td>
<td>7.00</td>
<td>1.71</td>
</tr>
<tr>
<td>Older sphagnum 1 ...</td>
<td>5.73</td>
<td>...</td>
<td>9.08</td>
<td>12.38</td>
<td>52.50</td>
<td>5.78</td>
<td>1.38</td>
</tr>
<tr>
<td>,, ,, 2</td>
<td>5.18</td>
<td>3.87</td>
<td>9.48</td>
<td>9.63</td>
<td>53.30</td>
<td>6.20</td>
<td>1.13</td>
</tr>
<tr>
<td>,, ,, 3</td>
<td>4.28</td>
<td>3.52</td>
<td>10.86</td>
<td>13.74</td>
<td>45.21</td>
<td>4.26</td>
<td>1.16</td>
</tr>
<tr>
<td>,, ,, 4</td>
<td>5.17</td>
<td>3.72</td>
<td>10.82</td>
<td>10.42</td>
<td>52.62</td>
<td>5.63</td>
<td>2.22</td>
</tr>
<tr>
<td>Woody peat ...</td>
<td>5.69</td>
<td>5.13</td>
<td>4.91</td>
<td>4.08</td>
<td>59.86</td>
<td>9.08</td>
<td>5.03</td>
</tr>
</tbody>
</table>

Peats contain between 10 and 20% of this constituent. The transition to low-moor is shown by the fall in cellulose to 4.08% in the woody peat.

Thirdly, the crude protein content of the low-moor peats, excluding the bottom mud, varies from 18.72 to 22.48%, whilst in the high-moor the range is from 4.26 to 7.00%, the woody peat indicating the transition with 9.08% of crude protein. The highest figure in the high-moor peat, 7.00%, is found in the heath peat, the so-called “grenzhorizont,” which is a constant feature of continental high-moors and may indicate a dry period.

W. L. Davies, in a comparison of the nitrogenous matter of peats, found evidence of more rapid protein degradation under fen conditions than in upland peat or high-moor conditions.

The complete succession from lowland peat, through forest peat, to upland peat can only be observed in basins or depressions formerly occupied by lakes. In other cases, development may begin at the forest peat or the upland peat stage.
CLIMATIC MOOR AND MOUNTAIN PEAT

There is a type of peat, termed by continental writers "Hangmoor," which is of common occurrence in mountainous regions. It is formed along slopes down which water is continually percolating from springs in higher ground. The character of the peat will depend on the nature of the percolating waters. Where these are poor in dissolved material, for example, surface water, or waters from highly acid rocks, the peat will be of the high-moor or sphagnum type. With richer waters, on the other hand, peats of the fen type may be developed, whilst in some cases trees may be established. Extensive deposits of acid peat of the upland type are found directly on boulder clay or colluvial material throughout the mountainous regions of Great Britain.

G. K. Fraser proposes the term climatic moor for those types of peat whose development is not primarily dependent on topography but on climate. In very humid regions such as the mountainous regions of Scotland and Wales, the stable soil type is of a moorland (peat) character in which the tendency is for impervious waterlogged layers of peat to develop. He distinguishes three sub-groups, namely, (a) Scirpus moor. North-western type with evenly distributed rainfall and low summer temperature; (b) Calluna moor or Calluna-Eriophorum moor. North-central Scottish type with drier summer and higher evaporation; (c) Molinia moor. Southerly type. Warmer and drier summer and higher evaporation.

HEATH AND FOREST PEAT

These terms are sometimes applied to the accumulations of humus on the surface of heath and forest soils. As the term peat is now generally restricted to humic material developed under wet anaerobic conditions, it is better to abandon the use of the term for humus accumulations developed under aerobic conditions with strong leaching. For such accumulations the term raw humus is better. (Cf. p. 204.)
BIBLIOGRAPHY


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CHAPTER XIII

SALINE, ALKALINE, AND SOLOTI SOILS

CONDITIONS OF OCCURRENCE

The soils to be described in the present chapter occur most commonly, but not exclusively, under arid climates. They owe their distinctive characters either to the presence of an actual excess of sodium salts, or to the predominance of sodium among the exchangeable bases—the latter a consequence of the former presence in the soil of sodium salts. In some cases, potassium salts may be present in appreciable amounts, but it is unusual for soils to occur in which potassium is the dominant exchangeable base.

The most commonly occurring groups are the saline, white alkali or solontshak* soils, and the alkaline, black alkali, or solonetz soils. To these must be added a third group, which has been brought to the notice of soil students by the Russian pedologists, namely, the soloti† or degraded alkali soils.

Before entering on a description of some representative types of these groups, it will be necessary to enquire into the conditions under which saline, alkaline, and soloti soils occur.

The first pre-requisite for their occurrence is the presence of sodium salts in the soil or the parent material. These originate usually from underground water in situations where the water-table is relatively near the surface. Such conditions occur generally in areas of basin-shaped topography. The salts may have originated directly from

*Also termed solonchak or solonchak.
†Also termed solodi or soloth.
the chemical weathering of silicates, resulting in the accumulation in the underground water of the soluble products of hydrolysis, or they may represent the vestiges of former seas or salt lakes. Stated generally, the most usual mode of origin is from saline telluric waters either in the vicinity of inland seas or salt lakes, or in depressions where the water-table is at or near the surface. In such cases there has usually been a change in the hydrological conditions. In primary saline soils, there has been a lowering of the water-surface whereby soils or their antecedent parent materials have emerged. In this way, for example, the saline soils of the Great Basin in the United States have originated, for they occupy an area formerly covered by the ancient Lake Bonneville, represented to-day by the Great Salt Lake. The fall in level has doubtless been accompanied by a concentration of soluble salts in the remaining water. A similar fall in water level is postulated also in the Caspian Basin.

In secondary or regraded saline soils, the salinization has taken place by a rise in the level of salt-bearing ground-water. This may have taken place naturally, as, possibly, in parts of Hungary; but more frequently it is encountered as a result of human interference. W. W. Mackie\(^1\) records that in 1873 the water-table at Fresno, in the San Joaquin valley of California, was 65 feet below the surface and the soil was practically free of salts. As a result of irrigation without adequate under-drainage, the level of the underground water in 1888 was only 2-3 feet below the surface and the soil was definitely affected with "alkali." Similar instances can be found in most irrigated areas where large tracts of formerly fertile soil have been ruined through the rise of ground-water.

We may class with the secondary saline and alkaline soils, those which, without any marked change in water-table, have undergone change through the use of irrigation waters in which sodium is the predominant cation. The continued use of such waters leads to a gradual replacement of exchangeable calcium, by sodium, so that even although
no considerable proportion of sodium salts is present, the soil becomes a sodium soil. In arid areas near to the ocean salinization may occur through the addition of sodium chloride brought down in the rainfall. L. J. H. Teakle estimates that the annual contribution of "cyclic" salt in W. Australian maritime districts may amount to 300 lb. per acre. In this connexion the existence of more pluvial conditions in past ages may be of importance. M. Gračanin attributes the salinity of certain soils in Adriatic islands to wind-borne salt.

SALINE SOILS

These soils are the white alkali soils of the earlier American writers, and the solonitchak soils of the Russian school. They contain an excess of sodium salts, generally the chloride or sulphate, and have a flocculated structure. They occur usually in depressions, and the distribution of salts varies with the season. During drought, they show white efflorescences of sodium chloride or sulphate, from which the American term white alkali is derived. During rain, the salts deposited at the surface are dissolved and temporarily washed down towards the water-table. A considerable variety of possibilities presents itself; for the character of the soil will depend on the position of the water-table, the amount and distribution of the rainfall, the concentration and composition of the soluble salts, and the general character of the soil of the region. Saline soils occur most commonly in arid, semi-arid, and semi-humid regions and may represent modified chernozems, chestnut earths, or grey earths. Where the water-table lies so near to the surface as to maintain permanently moist conditions, saline peaty or saline meadow soils may result.

Russia. Saline soils are generally described by Russian writers as structureless in contradistinction to the structural alkaline (solonetz) soils. A typical profile from the vicinity of Akmolinsk is described by G. Tumin:—
A. 0-1 cm. Light grey salt crust with carbonates, passing to a dark grey looser structureless horizon. Gradual transition to

A₂. Brown with grey streaks and mottlings. Weak carbonate reaction. Numerous small flecks of salt to 30 cm. From 30-45 cm., larger flecks.

C. Light brown salt bearing loam, containing carbonates.

The whole profile is moist and the saline ground-water is encountered at 110 cm.

Glinka⁵ describes a saline soil with affinities to the chestnut earths in the Jenisseisk province of Russia. The profile is as follows:—

0-1.5 cm. Loose dusty soil with plant roots.
1.5-9 cm. Rather more compact chestnut soil without marked structure.
9-19 cm. Lighter coloured than above horizon but especially porous.
19-44 cm. Lighter coloured and porous.

Finland. Saline soils may occur in humid regions where ground-waters containing salts may occur. B. Aarnio⁵ describes certain Finnish soils in which saline ground-water occurs. The following table shows the composition of the salts in different horizons of a typical profile consisting of:—

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Composition</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-10</td>
<td>Dark brown peat.</td>
</tr>
<tr>
<td>10-20</td>
<td>Fine sand.</td>
</tr>
<tr>
<td>&gt;20</td>
<td>Litorina clay.</td>
</tr>
</tbody>
</table>

At 15 cm., there is a gleys horizon showing brown streaks in the sand, whilst the clay is coloured yellowish-brown. The reaction of the surface horizon is less than pH 4. Removal of salts by drainage involves a lowering of the water-table, which is more than 1 metre below the surface. Capillary rise of salts is inappreciable, a circumstance which is connected with the humid climate. Saline water at this depth in an arid climate would dominate the character of the soil.

Saline soils, since they often occur in situations where the water-table is comparatively near the surface, may be expected to show affinities with the meadow soils. This is
Table XXXVI.—Salts in Vegetation-free saline soil, Isokyro, Finland (Aarnio)

<table>
<thead>
<tr>
<th>Depth</th>
<th>0-10 cm.</th>
<th>10-20 cm.</th>
<th>25-35 cm.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Percentage Water-soluble Salts</td>
<td>2.45</td>
<td>0.20</td>
<td>0.22</td>
</tr>
<tr>
<td>SiO₂</td>
<td>0.29</td>
<td>—</td>
<td>12.26</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>10.77</td>
<td>7.32</td>
<td>3.66</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>0.31</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>CaO</td>
<td>23.77</td>
<td>13.24</td>
<td>12.44</td>
</tr>
<tr>
<td>MgO</td>
<td>1.65</td>
<td>6.92</td>
<td>10.89</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.41</td>
<td>—</td>
<td>1.46</td>
</tr>
<tr>
<td>Na₂O</td>
<td>4.16</td>
<td>16.42</td>
<td>7.32</td>
</tr>
<tr>
<td>SO₃</td>
<td>56.78</td>
<td>48.84</td>
<td>48.76</td>
</tr>
<tr>
<td>Cl</td>
<td>0.85</td>
<td>7.12</td>
<td>3.20</td>
</tr>
<tr>
<td>Total</td>
<td>99.99</td>
<td>98.86</td>
<td>99.99</td>
</tr>
</tbody>
</table>

shown in some profiles by the presence of rusty flecks or mottlings in the lower horizons.

In extreme cases, saline soils are devoid of vegetation, as for example, in the vicinity of the Great Salt Lake, and the Dead Sea. In other cases, there is an open association of halophytic plants such as salt bush. With lower contents of soluble salts there is a transition to the normal soil of the region, which may nevertheless exhibit the characteristic structural and chemical features of the saline group.

**ALKALINE SOILS**

These soils, which are closely associated with saline soils, are termed black alkali by the American workers and solonetz by the Russians. They are characterized by the presence of sodium carbonate. Whilst the saline soils are in a state of flocculation, the alkaline soils have a strongly alkaline reaction and are deflocculated. The presence of alkaline soils in an area of saline soils is marked by depressions corresponding with the decrease in pore-space implied by the change of structure. In these depressions,
during rain, alkaline solutions of humic matter accumulate and, on drying, leave black deposits which give to these soils the American name "black alkali." The development of solonetz from solontshak is termed solonization.

The alkaline soils are further distinguished by the development of a structure profile. Typically, this consists of a laminated superficial horizon overlying a deep layer showing a columnar (rounded tops) or prismatic structure. The rounded tops of the columnar elements and the vertical sides frequently carry a white mealy coating of salts. The soil colours are grey or greyish-brown throughout.

The "structure" of alkaline soils is a consequence of their deflocculated state. When wet, the whole body of the profile consists of structureless material in the single grain state. On drying, shrinkage occurs and prismatic or columnar structures develop.

_Ukraine._ The following data are cited by D. G. Vilensky\(^7\) for a profile in the Dnieper region:—

0-19 cm. Light grey, platy, fine sandy eluvial horizon. The surface of the plates is powdered.

19-29 cm. Illuvial, dark grey, columnar horizon (columns 8-11 cm.). The columns have rounded tops with a meal-like efflorescence which also appears in places on the vertical sides of the columns.

20-48 cm. Transitional prismatic.

48-78 cm. Accumulation of sulphates. Light loess-loam with minute inter-layers of gypsum.

Another profile described by Vilensky showed the following characteristics:—

0-1.5 cm. Humus horizon. Semi-peaty, grey-brown loam.

1.5-10 cm. Dark brown short columnar. Columns are 4 cm. in diameter, clayey, compact, and do not effervesce with acid.

10-27 cm. Rather lighter coloured, breaking into large pillars, heavy loam.

27-83 cm. Dark brown with grey spots of effervescent salts. Compact loamy.

83-100 cm. Muddy brown with very numerous white spots. Loose non-structural.

100-120 cm. Bluish-grey flowing sand with water at 120 cm. under pressure.
The above profile contains chloride and sulphate in the upper horizons.

Hungary. The Hungarian alkali soils (Szík lands) contain very small proportions (0.15-0.20%) of soluble salts. They exhibit impervious horizons, considered by A. A. J. De Sigmund to the formation of such soils. The contrast between alkali and non-alkali soils in the character of the exchangeable cations is shown by the data in Table XXXVII.

**Table XXXVII. — Exchangeable Bases in Alkaline and Non-Alkaline Soils (De Sigmund)**

<table>
<thead>
<tr>
<th>Soil</th>
<th>CaO</th>
<th>MgO</th>
<th>K₂O</th>
<th>Na₂O</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bad Alkali</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hortobágy 1</td>
<td>2.60</td>
<td>1.34</td>
<td>0.94</td>
<td>9.62</td>
<td>14.50</td>
</tr>
<tr>
<td>Békéscsaba 5</td>
<td>8.32</td>
<td>3.86</td>
<td>0.14</td>
<td>7.30</td>
<td>19.62</td>
</tr>
<tr>
<td>Mezhohegyes</td>
<td>17.65</td>
<td>2.54</td>
<td>0.55</td>
<td>4.68</td>
<td>25.42</td>
</tr>
<tr>
<td>Non-alkali</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Keszthely 1</td>
<td>15.96</td>
<td>1.75</td>
<td>0.33</td>
<td>0.19</td>
<td>18.23</td>
</tr>
<tr>
<td>Keszthely 22</td>
<td>15.18</td>
<td>1.82</td>
<td>0.41</td>
<td>0.12</td>
<td>17.53</td>
</tr>
</tbody>
</table>

Whilst, in the bad alkali soils, the proportions of Na₂O to the total bases are 66.35, 37.21, and 18.41%, respectively, in the non-alkali soils, the proportions are 1.01 and 0.41%, respectively.

United States. Solonetz or alkali soils are widely distributed throughout the arid regions of the west. The following is a description of a western N. Dakota solonetz by C. E. Kellogg. The natural vegetation is western wheat grass, *Agropyron smithii*.

A₁. 0-1½ in. Dark greyish-brown clay loam thoroughly permeated with roots.
A₂. 1½-3 in. Very dark greyish-brown clay with hard coarsely granular structure.
B₂. 3-14 in. Brownish-black clay with very hard prismatic structure. Prisms from 1 to 6 in. in diameter, breaking with difficulty into irregular sharply angular pieces. Upper part faintly mottled rusty-brown.
B. 14-24 in. Very dark grey clay. Prisms much less distinct and soil more massive.
C. 24 in. +. Very dark grey massive clay soil; digs out in large massive chunks with difficulty. Only a few vertical cracks extend into this material.

Some mechanical mixing of material occurs through surface soil falling into vertical cracks. Effervescence with HCl found below 30 in.

RELATIONSHIPS OF SALINE TO ALKALINE SOILS

The relationship of alkaline to saline soils is now intelligible owing to the investigations of Gedroiz, De Sigmoid, Kelly, and others, on the base exchange reactions of these soils. The presence of an excess of sodium salts in saline soils results in the partial or complete replacement of their exchangeable calcium by exchangeable sodium. So long as excess of sodium salts is present they remain in the flocculated condition with neutral or slightly alkaline reaction. With removal of excess of sodium, the reaction becomes markedly alkaline and deflocculation occurs. The process has already been discussed in Chapter VI.

There is no hard and fast line of division between saline and alkaline soils, and all stages in the transition between the two types may be encountered. This, in addition to the great variety of possibilities in the conditions antecedent to salinization, results in very great variation in the character of these soils. The principal factors are, however, the position of the water-table, the composition of the groundwater, the nature of the soil and the degree of leaching, either naturally by rainfall or artificially by irrigation.

The problem of alkaline and saline soils is of the greatest importance for irrigated agriculture. Irrigation is usually practised in valleys and, unless special measures are taken to ensure drainage, the ground-water level rises, bringing sodium salts within the effective capillary range of the surface horizons. This leads in time to the replacement of calcium by sodium in the absorbing complex and, in addition, may involve such an increase in salt concentration in
the soil moisture as to inhibit plant growth. By the leaching action of irrigation water, the saline soil may become changed into an alkaline soil with deflocculation and increase in alkalinity.

In some cases the salinization may result from the irrigation water itself having excess of sodium over calcium ions, as is so frequently the case in S. Africa. The soil becomes changed from the calcium to the sodium soil with consequent deterioration in fertility.

It should be remarked that, although the presence of sodium carbonate is a characteristic of alkaline soils, the development of their distinctive properties is not dependent on the occurrence of large proportions of this salt. If, by the continued action of sodium salts, a soil is brought to the state in which the exchangeable calcium is replaced by sodium, the resulting sodium soil is potentially an alkaline soil and becomes deflocculated as soon as the concentration of the excess salts falls sufficiently low to permit the hydrolysis of the sodium clay. Under these conditions, sodium carbonate may be formed either by the reaction of the sodium clay with calcium carbonate, or by the action of carbon dioxide on the sodium hydroxide liberated by hydrolysis. These reactions may be represented by the equations:

\[
\begin{align*}
(1) & \quad \text{Na-clay} + \text{H}_2\text{O} = \text{NaOH} + \text{H-clay} \\
(2) & \quad 2\text{NaOH} + \text{CO}_2 = \text{Na}_2\text{CO}_3 + \text{H}_2\text{O} \\
(3) & \quad 2\text{Na-clay} + \text{CaCO}_3 = \text{Na}_2\text{CO}_3 + \text{Ca-clay}
\end{align*}
\]

Sodium carbonate should thus be regarded as one of the consequences of alkalinization rather than as the cause of the distinctive characters of alkaline soils. The percentage of sodium carbonate is in itself no measure of the degree of alkalinization of a soil. If carbon dioxide be excluded, it is possible to prepare completely deflocculated soils from which sodium carbonate is entirely absent.

An example of the artificial alkalinization of a soil by the action of sodium salts is seen in certain manurial plots at the Rothamsted station which have received repeated heavy
dressings of sodium nitrate. The soil has now become sticky and impervious, and exhibits some typical features of an alkaline soil.

D. S. Gracie, M. Rizk, A. Moukhtar, and A. H. I. Mustapha\(^{10}\) describe a type of akalinization in which biological activity plays a part. In certain Egyptian soils containing gypsum the anaerobic condition consequent on a high water-table results in reduction of calcium sulphate and formation of ferrous sulphide. This is accompanied by an increase in alkalinity which depresses the solubility of calcium and magnesium compounds and facilitates the formation of the sodium soil. At the same time calcium carbonate and magnesium silicate are precipitated.

Russian workers have attached first importance to morphology in the diagnosis of solontshak, solonetz and related soils. The term solonetz, as first used, denotes a soil having a certain type of structure profile. This was found later to be associated with the chemical properties which have been described above. C. F. Shaw and W. P. Kelley\(^ {11}\) have directed attention to the fact that soils exist which have the solonetz structure but lack the chemical characteristics associated with solonetz. N. I. Usov\(^ {12}\) relates the occurrence of the physical properties associated with magnesium solonetz to the ratio of exchangeable calcium to exchangeable magnesium. With Ca:Mg \(>3\), physical properties are dominated by calcium; with Ca:Mg \(<2\), the soil is heavier and shows a coarser structure. Where more than 40% of the base exchange capacity is represented by magnesium the solonetzic properties are well developed, but to a smaller degree than in the case of sodium soils. J. H. Ellis and O. G. Caldwell\(^ {13}\) also have described soils with solonetz structure containing little or no exchangeable sodium. It seems clear, therefore, that if the term solonetz is to be retained in its original sense, chemical composition must be assigned a subsidiary rôle. The writer would prefer to abandon the term as a description of a world group and use the term alkaline soil, which has a definite chemical significance.
[Photo. C. E. Kellogg.

SOLODIZED-SOLONETZ, N. DAKOTA.

PLATE VIII.]
SALINE, ALKALINE, AND SOLOTI SOILS

SOLOTI OR DEGRADED ALKALI SOILS

The hydrolysis of sodium soil with production of sodium hydroxide involves also the formation of the hydrogen soil, as shown above. In the presence of excess of calcium carbonate, calcium soil is speedily re-formed, if indeed it is not formed by the direct inter-action of the sodium soil and calcium carbonate. In the absence of calcium carbonate, however, degradation of the hydrogen soil into silicic acid and sesquioxides occurs. The sesquioxides become leached out, leaving a bleached eluviated horizon relatively rich in silica. The process is in its results similar to podsolization, but the actual degree of destruction is greater. The A horizon of soloti soils may contain as much as 12% of free silica of secondary origin. The formation of soloti from solonetz is termed solodization or solotization.

A typical Russian soloti profile is described in the following terms by Korotky\textsuperscript{14}:

A\textsubscript{6}. Peaty layer with grass cover; 3-5 cm.
A\textsubscript{1}. Grey or ashen-grey layer with tongue-like projections from the peat layer; 1-10 cm.
A\textsubscript{2}. Whitish grey, with numerous rust spots, and ferruginous concretions; typical podsolic laminated structure; 10-15 cm.
B\textsubscript{1}. Brown varying to rusty or bluish-grey (humus); abundant humous mottlings; a sticky clay; numerous "ortstein" grains in the upper portion, forming in the lower portion a black or dark brown material (iron and manganese compounds); on drying breaks into prismatic lumps.

United States. C. E. Kellogg\textsuperscript{7} describes solodized solonetz soils in western N. Dakota. The following is an example. It is from the same region as the solonetz described on p. 385. A similar profile is shown in Plate VIII.

A\textsubscript{1}. 0-1 in. Brownish-grey loam with soft crumb or mulch-like structure.
A\textsubscript{2}. 1-3 in. Greyish-brown loam, platy and friable.
B\textsubscript{1}. 3-5 in. Dull greyish-brown silt loam. Relics of prismatic structure but distinctly platy. Easily friable to medium crumb structure.
B\textsubscript{2}. 5-8 in. Dark-brown clay with well-developed hard columnar structure. Cups of columns well developed.
B,. 8-12 in. Olive brown clay with irregular prismatic structure. Easily crumbling to sub-angular difficulty friable nut-sized pieces.

B1. 12-22 in. Olive-brown clay, similar to above, except that carbonates are present. Numerous white flecks and streaks.

C,. 22-26 in. Olive-grey sandy clay, highly mottled with white. Compact in places and digs out in brittle easily friable pieces.

C. 26 in. +. Olive-grey sandy clay till, mottled white. Compact in places as above, but pieces become very hard on drying.

Saline, alkaline, and solotı soils occur commonly in close association often as complexes differentiated by relief forms and local hydrological conditions. C. E. Kellogg has studied such a complex association in western N. Dakota. The normal sequence of development is chestnut-brown earth→solonshak→solonet→solotı→eroded solotı. With an erosion of the solotı profile, vegetation is re-established in the exposed B horizon and the normal profile may be re-constituted. In some cases regeneration may take place without intervening erosion. The distribution is exceedingly complex and a pronounced micro-relief develops through differential erosion of the eluviated horizons of the solonetz-solotı complex. A distinction is drawn between salinization by flooding and salinization by capillary action.

Saline and related soils occur most commonly in arid regions and are therefore associated most usually with pedocalic soils, where their occurrence is generally determined by topographical factors. They are extensively found in such areas as the Great Basin of the United States, the Caspian region of Russia, and the interior of Australia. They come most frequently to the notice of soil investigators through their development in irrigated tracts. It is a common experience in such regions that soils deteriorate under irrigation. This may be due either to the rise of a saline water-table or to the use of water containing excess of sodium salts. In many irrigated areas there are considerable tracts which have gone completely out of cultivation through akalinization.
Solodized-solonetz landscape, N. Dakota.

Photo. C. E. Kellogg.

Plate IX. [To face page 390.]
Whilst it is known that derelict alkali lands can be reclaimed by such methods as dressing with gypsum, in order to reconstitute the calcium soil, comparatively little has been done in practice to check deterioration in irrigated soils. The reclamation of such lands and the prevention of deterioration in land now under irrigated cultivation forms one of the most important tasks of applied pedology.

The reclamation of tidal or submerged lands involves the change of saline into normal soils. In Holland, the home of such works, the transformation is facilitated by the humid climate and by the presence of calcium carbonate in the soil.

BIBLIOGRAPHY

5 Glinka, K. D., ibid.
CHAPTER XIV

SOILS OF THE HUMID TROPICS AND SUB-TROPICS

GENERAL CHARACTERISTICS

The devotion of a separate chapter to the discussion of tropical soils must not be taken to imply that they can be considered as in any way constituting a great world group. In fact, representatives of most of the recognized world groups can occur in the tropics, but the special conditions occurring in those regions, above all the high temperatures, result in certain distinctive soil characters. It is, therefore, convenient, if only for descriptive purposes, to consider them together. Certain soils occurring in the tropics, but more thoroughly studied in temperate regions, have already been mentioned. In the present chapter, therefore, we shall not deal with tropical pedocals, principally because in the present state of knowledge they are more conveniently considered along with the better known pedocals of temperate climates. If the increase of our knowledge of these soils made possible a fuller treatment, it might prove better to consider them together with the other soils of tropical regions. In the present chapter, therefore, we shall discuss only the soils of the humid tropics.

The most important circumstance differentiating the soils of the tropics from other soils is, of course, the temperature conditions under which they are developed. The mean temperature within the tropics is generally above 25°C, and in latitude 30°, is about 20°C. Bearing in mind that all chemical changes increase in velocity with rise of temperature, a rise of 10°C approximately corresponding with a
doubling of velocity, it is easy to understand that the hydrolysis of mineral silicates, which forms the most important process in chemical weathering, proceeds with much greater rapidity under tropical conditions.

In addition, there has been an absence, during recent geological time, of glacial interference. Tropical weathering has thus operated over a much longer period of time than weathering in regions such as Northern Europe, where it can only date from the close of the last glacial epoch. There is also no interruption of chemical action by winter, as in temperate and cool regions. It is not surprising, therefore, that the depth to which epigene weathering extends is considerably greater than in temperate and cool climates. It is not uncommon to find fifty feet or more of weathered material.

The A, B, and C nomenclature used in the description of soils of temperate regions has not been used so frequently by students of tropical soils. Indeed, in many cases considerable uncertainty may exist as to the depth of the solum, for there may be little apparent change below the top few inches. It might be justifiable to regard soils developed over deep layers of weathered material as secondary (see p. 65) in that the processes that have formed them from the parent rock date back in many cases through geological time, whereas soil profile development is mainly contemporary. Many of the distinctive characters of tropical soils are, indeed, attributable rather to the parent material than to the operation of present-day processes of profile development.

The parent material of tropical soils, being the product of age-long processes, is often in an advanced stage of weathering. Since hydrolysis of rock-forming silicates involves loss of silicic acid, alkalis, and alkaline earths, the parent material will tend to be markedly sesquioxide in character, in other words to be ferrallitic rather than siallitic.

Although the general tendency is towards a ferrallitic product, the actual course of weathering may vary. The most typical case is where the first product of weathering,
adjacent to the decomposition rock is of a siallitic (kaolinitic) type, whilst the overlying horizons of weathered material are ferrallitic or, according to the nomenclature of certain writers, lateritic. On the other hand, F. J. Martin and H. C. Doyne describe the weathering of norite in Sierra Leone, in which the first product appears to be ferrallitic. J. B. Harrison also regards the primary process as resulting in a ferrallitic (lateritic) product. The difference may be attributable to the parent rock in that the primary product is generally siallitic in the case of acid rocks and ferrallitic in the case of basic rocks.

One circumstance, possibly fortuitous, that tends to colour our views on tropical soils is that so much of the published data relate to soils developed in material derived from the weathering of igneous rocks. On the other hand much of the best known work on the soil of temperate regions, including most of the Russian and Central European work, relates to soils derived from secondary materials such as loess, glacial drift, and alluvium. It is possible that some, at any rate, of the characteristics commonly associated with tropical soils may be attributable to the type of material from which they have been developed, rather than to the particular conditions of profile formation.

Tropical soils often consist almost entirely of the products of chemical weathering, together with a greater or less admixture of unweatherable minerals such as quartz, magnetite, etc. Weatherable minerals such as felspars and ferromagnesian minerals, are, in some cases, almost entirely absent. Such soils, if derived from basic rocks that do not contain quartz, may contain very high proportions of clay, as much as 80% being by no means uncommon.

Coming now to the actual processes of profile development we have, as in the case of weathering, the effect of the high temperatures, which not only favour plant growth but also those chemical and biological processes whereby plant residues undergo oxidative decomposition. We shall return to this later.
Of equal importance with temperature in determining the type of profile development in the humid tropics is the character of the rainfall. Equatorial regions have generally high rainfalls, but two peculiarities must be noticed.

Firstly, the rainfall tends to fall in heavy downpours, exceeding in intensity the heaviest rain encountered in higher latitudes. There are instances of as much as 6 inches of rain falling in an hour. Such intensity of rainfall can cause a strong erosion, particularly where the natural vegetative cover has been removed as a preliminary to cultivation. This leads to the development of extensive alluvial and colluvial fans in valleys. A further consequence of the prevalence of erosion is the production of truncated or immature profiles. Many of the difficulties which present themselves in the study of tropical soils might be obviated if the part played by erosion were more clearly appreciated.

Secondly, except in the regions of tropical rain forest, where rainfall prevails throughout the year, there is a general tendency for tropical humid climates to be divided into well marked wet and dry seasons. During the wet season, leaching by percolating waters may occur, resulting in a general impoverishment in bases and the production of acid profiles with consequent instability of the clay complex. The conditions during the dry seasons are totally different. By reason of the high temperature, evaporation is very intense and the soil becomes rapidly dried out to a considerable depth. Such climates are sometimes termed "monsoon" climates.

The prevailing colour of freely drained tropical soils is red, the actual tinge varying somewhat according to the amount of organic matter present and the mutual effect of the parent material and the pedogenic processes. In basin-shaped areas with impeded drainage, and also where the parent material has a high lime status, dark-coloured or grey soils may result. It may be assumed that both the processes affecting the inorganic weathering complex and the decomposition of plant residues proceed differently according
as the soil profile is developed with free or restricted drainage.

Red soils appear, then, to result under conditions of free drainage and low lime status. Contrary to the impression produced by their bright red colour, such soils, until they are cultivated, may contain appreciable proportions of organic matter, particularly where the natural vegetation forms a close association such as forest, forest steppe, or savannah.

Topography has a marked effect on profile development. Whilst upland areas of convex topography are strongly leached and more or less eroded, lowland areas with concave topography are permanently or seasonally wet and carry soils having affinities with the gleys soils of temperate climates. This often results in a characteristic pattern in which the elevated areas are occupied by freely drained red soils and the depressions by grey or black soils with impeded drainage.

Soils subjected to submergence undergo characteristic changes, provided sufficient organic matter be present. Under these conditions, anaerobic decomposition occurs and rapid losses may ensue. There is also an increase in the solubility of the iron, manganese, calcium and magnesium, which are changed to hydrogen carbonates, the iron being in the ferrous state. The reductive processes result in a change from the red colour of well aerated soils to the grey colour associated with bottom soils. Where water movements can occur, the dissolved bases may be lost from the soil with consequent impoverishment.

The rôle of organic matter in tropical soil formation has been discussed by E. C. J. Mohr⁴. The amount of organic matter in a soil represents the balance between the addition of material in the form of plant residues and the destructive oxidation of added material by micro-organisms. The general relationship between plant growth and temperature may be expressed by a curve having its minimum at about 0°C, its optimum at 25°C and its maximum at something over 40°C. Given adequate moisture conditions, we may
expect the maximum production of organic matter and, therefore, the maximum addition of plant residues at about 25°C. The curve connecting micro-biological activity with temperature shows a different course. The minimum lies somewhat above that for the growth of the macro-flora. Mohr places it at about 10°C, but this is probably too high. The optimum lies at about 35-40°C, whilst the maximum may lie at 70-80°C, well above the temperatures encountered in actual soils. Mohr distinguishes micro-biological decomposition under aerobic and anaerobic conditions, the intensity of decomposition being considerably greater under aerobic conditions.

Considering, now, the course of accession and destruction of organic matter in aerobic soils, the positive balance between the two processes increases up to about 20°C, after which the rate of destructive microbiological action increases more rapidly than plant growth with rise in temperature, until at somewhat above 30°C, destruction is sufficiently rapid to keep pace with additions of organic matter.

Under anaerobic conditions, less intense decomposition leads to the maximum balance between accession and destruction occurring at a higher temperature. The point at which gains are balanced by decomposition lies at about 35°C.

From the above considerations, it might be expected that virgin tropical soils with good aeration would have a very low organic matter content. This is by no means always the case. In the examination for another purpose of a range of soils, which included many samples from E. Africa and other parts of the tropics by G. W. Robinson and M. Richardson, organic carbon contents were determined and the figures obtained were not markedly lower than those to be expected from a similar textural range under British conditions. Further investigation is necessary, but it may be tentatively suggested that, although plant residues undergo rapid decomposition under tropical conditions, the actual humus is more resistant. The recurrent drought occurring
under monsoon climates may be conservative of soil organic matter as in the case of the chernozems.

The humus of tropical red soils is generally paler in colour than that of soils in temperate regions. G. Milne reports that soils from Tanganyika with about 4.1% organic carbon are not markedly humous in appearance. He considers that this humus has little crumb-forming effect.

Except in regions of continuous rain, the humid tropics generally have marked wet and dry seasons. During the dry seasons, the soil moisture becomes relatively concentrated and tends to move upwards to the surface where it evaporates. There will thus occur a precipitation of materials from solution, principally in the surface horizon, which thus becomes an horizon of accumulation. The precipitation which occurs will be to some extent irreversible and result in the formation of the concretions which are such a distinctive feature of tropical soils. On the side of increasing humidity, concretionary deposits will tend to become less frequent.

H. H. Bennett and R. V. Allison have analysed the concretions (perdigón) from some typical Cuban soils. They are often markedly sesquioxidic in character but always contain a certain amount of silica. In some cases the silica may amount to over 50%. Such concretions are similar in composition to fairly siliceous clays.

Tropical soils frequently show an increase in acidity, as measured by pH value, in descending the profile. This has been discussed by H. C. Doyne, with special reference to Nigerian soils.

RED LOAMS AND RED EARTHS

As has been mentioned above, mature soils developed in the tropics under conditions of free drainage are generally red in colour. Tropical red soils are often loosely termed lateritic. The term laterite has been used in many senses, and in the sequel we shall attempt to restrict its application to one particular type of profile to be discussed at a later
point in this chapter. The red colours of tropical soils may be attributed to the presence of free ferric oxide in one or more of its different forms. Some red soils may contain very high proportions of ferric oxide, but this cannot always be inferred from their colour. Free alumina is also present in notable proportions in many tropical soils. For soils with high proportions of sesquioxides the term allitic was proposed by A. Harrassowitz in contrast with soils having a more siliceous type of weathering complex which are termed siallitic. In the present account we shall use the term ferrallitic in place of allitic to denote soils containing high proportions of sesquioxides in the weathering complex. A completely weathered soil, i.e., a soil in which the weatherable minerals have undergone complete desilicification with a resulting weathering complex consisting only of sesquioxides may be termed ferrallite. We consider these terms preferable to lateritic and laterite, used by many earlier authors but open to objection by reason of their lack of precision. Finally, it would appear desirable to restrict the terms siallitic, allitic, and ferrallitic to the description of soils as materials rather than as profiles.

If it be assumed that weathering under humid tropical conditions can produce a residual complex with an excess of sesquioxides, variations may occur according to (a) the extent to which desilicification has occurred, and (b) the extent to which the soil or its individual horizons have received accessions of sesquioxides by illuviation. Apart from the most extreme ferrallites, which must be regarded as exceptional cases, tropical red soils always contain some silica in the clay complex and the presence of alumino-silicic clay constituents must be assumed.

The investigations hitherto made by X-ray methods on the clay complex have not disclosed any minerals having molecular $\text{SiO}_2/\text{R}_2\text{O}_3$ ratios less than 2.0. Whilst S. Mattson, by removal of excess sesquioxides, using hot aluminium chloride solution, obtained a residue of the halloysite type, the amounts of free sesquioxide found by
F. Hardy\textsuperscript{19} in tropical soils would imply the presence of minerals having $\text{SiO}_2/\text{R}_2\text{O}_3$ ratios less than 2.0.

P. Vageler\textsuperscript{11} draws a distinction between red loams and red earths. We have seen that the tendency of weathering under tropical conditions is towards the production of a weathering complex relatively enriched in sesquioxides by the removal of silicic acid or by accession of illuvial sesquioxides. In comparatively young soils these changes are not far advanced. The clay is of a siliceous type and the soil is described as a red loam. In the red earths, removal of silicic acid or accession of sesquioxides has proceeded sufficiently to give a weathering complex of a predominantly sesquioxidic character. A. Eichinger\textsuperscript{12} distinguishes three stages in soil development under humid tropical conditions, namely, younger red earth, older red earth, and laterite. All three types may occur under a uniform forest cover and the differences only become apparent after the institution of cultivation.

A red loam profile (Eichinger’s younger red earth) consists of a deep and uniform layer of fairly plastic clay, often with yellow mottlings in the lower horizons and a gradual transition through a zone of decomposition to the parent rock.

The distinction between red loams and red earths has been used by many writers, notably by the late G. Milne. Although red earths as materials represent a more advanced stage of chemical weathering than red loams, it should not be assumed that they are necessarily more mature as soils, as might seem to be implied by Eichinger’s nomenclature. The difference in development relates merely to the parent material. The soils of the south-eastern United States described as red podsolic soils would appear to be akin to, if not identical with, the tropical red loams. The redness is essentially a quality of the parent material, for the superficial horizons may show some bleaching.

\textit{Cuba.} The soils of the Limones series in Cuba,
described by H. H. Bennett and R. V. Allison\textsuperscript{13}, may be cited as examples of the red loams. The profile consists of:—

0-4 in. Reddish-brown to dull red clay; cracking into hard clods on drying.
4-12 (18) in. Pinkish-red clay, cracking in dry season. Fragments of parent rock present in lower parts of horizon.
12 (18)-24 in. Partly decomposed soft serpentine rock, greenish or bluish or hard rock.

The pH given is 6.72, the figure presumably referring to the surface soil. No data are available for the composition of the colloidal clay, but the silica and sesquioxide figures for the total soil of two successive layers are as follows:—

<table>
<thead>
<tr>
<th>Depth</th>
<th>SiO\textsubscript{2}</th>
<th>Al\textsubscript{2}O\textsubscript{3}</th>
<th>Fe\textsubscript{2}O\textsubscript{3}</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-16 in.</td>
<td>51.25</td>
<td>17.40</td>
<td>10.69</td>
</tr>
<tr>
<td>16-42 in.</td>
<td>45.85</td>
<td>13.62</td>
<td>7.84</td>
</tr>
</tbody>
</table>

The clay is thus in all probability of a siliceous type, as might be expected from its plastic and cohesive character. The relative shallowness of the profile and the absence of concretions suggest a soil in the early stages of development.

The Nipe clay, on the other hand, is an example of a red earth, in which development has proceeded almost to the stage of a ferrallite. It is thus described by Bennett and Allison\textsuperscript{14}:

0-26 in. Deep red friable clay with abundance of perdigón (concretions).
26-40 in. Red or brownish-red friable clay with less perdigón.
40 in.-13 ft. Yellow, exceedingly friable ochreous material.
13 ft.-16 ft. Reddish, yellowish, and almost black, exceedingly friable material extending to bed rock.
16 ft. + Serpentine, weathered in upper part, faced along fracture lines with thin coating of bluish to purplish black.

The pH varies about 6.0. The composition of the colloidal clay (G. Edginton) is:—

<table>
<thead>
<tr>
<th>SiO\textsubscript{2}</th>
<th>Al\textsubscript{2}O\textsubscript{3}</th>
<th>Fe\textsubscript{2}O\textsubscript{3}</th>
</tr>
</thead>
<tbody>
<tr>
<td>10.19</td>
<td>15.84</td>
<td>62.51</td>
</tr>
</tbody>
</table>
The highly sesquioxidic character of the clay is reflected in its friable character, whilst its maturity is indicated by the great depth of weathering. The exchangeable bases show the exceedingly low figures of 1.1 milligram equivalents per 100 g. of soil.

The Nipe clay is a highly weathered soil, but it is by no means certain that it represents simply a residual product. The abundant perdigón in the upper horizons certainly represents illuvial material, and it may be that the colloidal clay also is enriched by secondary accumulations of sesquioxides.

_Central America._ Friable clays of the red earth type are of frequent occurrence in Central America. H. H. Bennett describes a typical profile in the following terms:

1. Surface layer 1/2 to 3/4 in. in thickness in which organic matter has slightly darkened the red soil.
2. Red clay uniform to 2-8 feet.
3. Red clay streaked or mottled with material of yellowish, bluish-grey, and whitish or cream colours often mixed with partly decayed rock material.

These soils are devoid of carbonates and low in exchangeable bases. The average silica-sesquioxide ratio for the total soil varies from 0.15 to 1.96.

_E. Africa._ The writer has been fortunate, through the courtesy of the late Mr. G. Milne, of the Amani Institute, Tanganyika, in securing a number of typical red soils from East Africa. The analytical data obtained from them present many points of interest.

The first profile to be discussed is from Muhesa, Tanganyika. It occurs on the breast of a long slope at about 600 ft. altitude, under a rainfall of about 50 inches per annum. Mr. Milne's description of this profile, which is that of a typical red soil, is as follows:

"Reddish chocolate top soil, 0-5 inches; red clay below; quartz gravel layer at 30-36 inches; clay again below that, becoming yellow-brown; then yellow to rock-brash and then raw gneiss,"
The analytical data are shown in Table XXXVIII.

**Table XXXVIII.—Analytical data for Muheesa soil**

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Clay %</th>
<th>Organic Carbon</th>
<th>pH</th>
<th>Clay Fraction</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>SiO₂/Al₂O₃</td>
</tr>
<tr>
<td>0-5 in.</td>
<td>55.25</td>
<td>1.65</td>
<td>7.7</td>
<td>2.25</td>
</tr>
<tr>
<td>12-18 in.</td>
<td>70.94</td>
<td>0.31</td>
<td>6.6</td>
<td>2.23</td>
</tr>
</tbody>
</table>

A profile from Kwamshindi, Tanganyika, occurring at 2,700 feet with an annual rainfall of 80 inches, under secondary bush, formerly rain forest, is thus described:—

"Dark coloured chocolate top soil, 0-12 inches; light reddish-chocolate subsoil, cubical structure, much angular quartz gravel; passes through orange and drab-yellow (2 feet thick) to raw gneiss at about 12 feet."

The analytical data are shown in Table XXXIX.

**Table XXXIX.—Analytical data for Kwamshindi soil**

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Clay %</th>
<th>Organic Carbon</th>
<th>pH</th>
<th>Clay Fraction</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>SiO₂/Al₂O₃</td>
</tr>
<tr>
<td>0-12 in.</td>
<td>45.2</td>
<td>3.21</td>
<td>6.3</td>
<td>2.01</td>
</tr>
<tr>
<td>12-20 in.</td>
<td>45.05</td>
<td>1.89</td>
<td>6.0</td>
<td>1.95</td>
</tr>
</tbody>
</table>

Another profile from gneiss near Amani, at 3,300 feet, rainfall 85 inches, in virgin forest is described as

"Dark blackish-brown top soil, 0-8 inches: dark orange-brown subsoil, much rock-brash at 27 inches and below; raw gneiss probably at about 6-8 feet."

The analytical data are shown in Table XL.

**Table XL.—Analytical data for Amani soil**

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Clay %</th>
<th>Organic Carbon</th>
<th>pH</th>
<th>Clay Fraction</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>SiO₂/Al₂O₃</td>
</tr>
<tr>
<td>0-8 in.</td>
<td>27.4</td>
<td>6.17</td>
<td>6.1</td>
<td>1.33</td>
</tr>
<tr>
<td>12-18 in.</td>
<td>40.6</td>
<td>1.72</td>
<td>5.1</td>
<td>1.33</td>
</tr>
</tbody>
</table>
An important fact which emerges from these analyses and those of a large number of other soils from East Africa is that larger proportions of organic matter, as judged from the organic carbon, are present than would be thought possible under tropical conditions. The soil from virgin forest at Amani contains over 6% organic carbon, corresponding with about 10-11% of organic matter. Even under secondary bush, the Kwamshindi soil contains the respectable content of 3.21% organic carbon. It should be added, however, that these soils are mainly from fairly high altitudes.

The figures for the composition of the clay fraction in the first two soils do not indicate a pronounced lateritic development. They might be paralleled by many figures for moderately eroded soils in the uplands of Wales. The third example indicates a definitely lateritic type of weathering complex. But even here, the sesquioxidic character is mainly due to the high proportions of ferric oxide.

Southern Nigeria. H. C. Doyne and W. A. Watson describe a type of soil to which they give the native name ilepa. It is derived from acid igneous rocks under mean temperatures of 70-90°F and a seasonal rainfall of 50-60 in. per annum. The profile consists of the following horizons: (a) brown sandy soil to about 3 ft.; (b) an horizon of concretions, black, orange, and red in colour, sometimes forming a hard pan-like conglomerate; (c) a layer of red clay mixed up with and below the concretions, becoming white-mottled with increasing depth; (d) white clay mottled with red; and (e) rotting rock. The structure of the parent material is recognizable in the clay horizons. The profile is not markedly sesquioxidic. It is considered that the concretions are formed by deposition of hydrated ferric oxide. The surface soil may be adventitious. An interesting feature noticed by Doyne and Watson is the decrease of pH with depth. This has been noted in other tropical soil profiles.

Mauritius. N. Craig and P. Halais give an account of soils derived from doleritic basalt under a mean annual tem-
SOILS OF THE HUMID TROPICS

perature of 23° C and rainfalls varying from 25 to 150 in. per annum. They divide the soils into mature and immature lateritic soils. The mature soils are brownish-yellow, whilst the immature soils are red in colour. The mature soils show a negative correlation between clay content and rainfall. In both classes of soils there is a decrease of the silica/sesquioxide ratio of the clay with rainfall as may be seen from the following mean figures:—

<table>
<thead>
<tr>
<th>Rainfall in inches</th>
<th>Clay %</th>
<th>SiO₂/R₂O₃ of Clay</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mature</td>
<td>Immature</td>
</tr>
<tr>
<td>25-50</td>
<td>73</td>
<td>49</td>
</tr>
<tr>
<td>50-75</td>
<td>67</td>
<td>53</td>
</tr>
<tr>
<td>75-100</td>
<td>56</td>
<td>49</td>
</tr>
<tr>
<td>100-125</td>
<td>49</td>
<td>44</td>
</tr>
<tr>
<td>125-150</td>
<td>45</td>
<td>43</td>
</tr>
</tbody>
</table>

S.E. United States. Red soils, described in the American system as red podsolic soils, are common throughout the South-Eastern United States and may have affinities with the red soils of the tropics.

A profile near Greensboro, N. Carolina, described by R. C. Jurney¹⁷, showed the following characteristics:—

A. 0 to 9 in. Slightly reddish-brown clay loam.
B₁. 9 to 36 in. Deep red heavy brittle clay.
B₂. 36 to 60 in. Light red friable crumbly clay.
C. 60 in. + Ochreous yellow, black and reddish-brown decomposed diorite.

There is a marked accumulation of clay in the B₁ horizon. The organic matter decreases from the surface, and the pH falls from 6.3 in the A horizon, to 4.1 in the C horizon. The silica-sesquioxide ratios in the colloidal clay of the successive horizons are 1.46, 1.49, 1.42, and 1.40, respectively. Ferric oxide increases with depth, the figures being 15.71, 18.94, 24.13, and 23.64%, respectively.

L. D. Baver and G. D. Scarseth¹⁸ have discussed the genesis of red soils in Alabama, with special reference to soils of the Susquehanna series. They consider that the northern limit of the type of weathering characterized by
sesquioxide enrichment corresponds with the 61°F mean annual isotherm.

**Western Samoa.** F. T. Seelye, L. I. Grange and C. H. Davies describe ferrallitic soils developed under 110-200 in. mean annual rainfall and 78.79°F. mean annual temperature from weathered basalt. In extreme cases they consist almost entirely of hydrated ferric and titanium oxides. They show a "lithomarge" layer over the basalt, enriched by silica washed down from the upper horizons.

**Brazil.** The Brazilian "terra roxa," well known for its fertility, is not, according to F. W. Freise, a ferrallitic complex. The relative predominance of kaolinitic and ferrallitic weathering depends on the balance between rainfall and evaporation. The descriptions of V. Agafonoff are in agreement with these views, from which it may be concluded that these soils are of red loam or "younger red earth" character.

**Southern China.** Red soils in Southern China are described by C. F. Shaw. They occur under a mean annual rainfall of 50-60 inches. A typical profile under small herbs and grass consisted of 2-5 inches of brownish red clay loam, friable but with very little organic matter, over a clay subsoil of a red colour with reticulate yellow and grey mottling. The parent material is shale and sandstone.

**Indo-China.** V. Agafonoff reports red soils from Indo-China. They are derived from basaltic rocks, and the ferruginous crusts which are developed in certain cases are considered to result from upward translocation of sesquioxides during drought.

**FRIABLE AND NON-FRIABLE SOILS**

H. H. Bennett had drawn a distinction between friable and non-friable tropical soils. The former appear to belong to Vageler's red earths and Eichinger's older red earths and laterites, and the latter to Vageler's red loams and Eichinger's younger red earths. They are distinguished principally by the character of their weathering complex,
the non-friable soils being relatively siliceous, whilst the friable soils are relatively sesquioxidic in character.

Siliceous and sesquioxidic soils are strongly contrasted in their textural properties. In the former type, the weathering complex is definitely of a clay character and, where present in considerable amounts, gives rise to the characteristic plasticity and cohesiveness of clay soils. In soils whose clay complex contains excess of sesquioxides, the properties of plasticity and cohesiveness are less strongly marked. A soil may contain very high proportions of this weathering complex and yet be of an open friable character.

Comparing friable and non-friable soils in Central America, Bennett obtained the following data (Table XL1).

**Table XL1. — Average and range of silica, ferric oxide, alumina, total bases, and $\text{SiO}_2/\text{R}_2\text{O}_3$ in friable and non-friable soils (H. H. Bennett)**

<table>
<thead>
<tr>
<th></th>
<th>$\text{SiO}_2$</th>
<th>$\text{R}_2\text{O}_3$</th>
<th>$\text{R}_0 + \text{R}_3\text{O}$</th>
<th>$\text{SiO}_2/\text{R}_2\text{O}_3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Friable...</td>
<td>29.30(6.00-41.87)</td>
<td>49.19(31.19-66.72)</td>
<td>1.32(0.13-6.55)</td>
<td>1.25(0.15-1.96)</td>
</tr>
<tr>
<td>Non-friable</td>
<td>52.33(41.27-77.00)</td>
<td>39.81(12.18-33.78)</td>
<td>4.50(6.85-17.17)</td>
<td>3.71(2.00-11.60)</td>
</tr>
</tbody>
</table>

The soils of the humid tropics and, indeed, of the humid sub-tropics, appear to present all stages between laterite and soils in which free sesquioxides, even if present, do not dominate the properties of the soil, although their presence may be inferred from the red colours which generally obtain.

**Affinities of Tropical Red Soils with Brown Earths**

The tropical red soils hitherto described, with the possible exception of the red soils from Southern China, are derived by primary weathering from crystalline rocks. Much of the material on which current ideas of soil classification are based consists of studies on soils of secondary weathering, i.e., in which products of chemical weathering are already present in the parent material. It is therefore important to bear in mind the possibility that some of the
characteristic features of the soils of the humid tropics might be found in soils of primary weathering in temperate climates. Apart from the greater degree of weathering and the occurrence of concretionary material, the writer sees no essential difference between the reddish-brown soils derived from crystalline rocks in Britain and the younger red loams of East Africa. It must be remembered that, even in the humid tropics, concretionary material is not invariably present and, except in ground-water soils, becomes an important constituent only under climatic conditions marked by intense drought.

The more vivid red colours of tropical soils may be attributed to difference in the degree of hydration of the ferric oxide. Whilst, under temperate climates, the hydrates are of the limonitic type, the less hydrated turgite, goethite, and even haematite, are probably present in tropical soils.

YELLOW EARTHS

This is a rather ill-defined group of soils of which representatives may occur in the tropics but which is more commonly encountered in sub-tropical and warm temperate climates. The yellow colour is probably connected with a higher degree of hydration of the ferric oxide in these than in the red soils. Yellow earths probably form a transition group between the tropical red soils and the brown forest soils of temperate climates. Certain yellow soils, such as the Norfolk series, in the S.E. United States, are held by C. F. Marbut to differ from adjacent red soils on account of the removal of iron during an earlier stage of impeded or sluggish drainage.

Yellow soils may also result from incipient podsolization of red soils or may represent early stages in the development of red soils. They may also be formed by the weathering of parent materials of low iron content. The use of the term has been discussed by W. Hollstein\textsuperscript{24}. In view of the uncertain position of these soils, it would appear inadvisable
to regard them as constituting a world group. Yellow soils can obviously originate in a variety of ways.

The recognition by the United States Soil Survey of a group known as yellow podsolic soils may provide a convenient repository for the soils hitherto described as yellow earths. If, as in the case of the red loams and red podsolic soils we regard the characteristic colour as belonging to the parent material, we can group together the grey-brown podsolic soils (brown earths), yellow podsolic soils (yellow earths), and red podsolic soils (red loams) as a sequence reflecting the effect of increasing temperature.

**LATERITE**

In 1807, H. Buchanan described a type of red soil occurring frequently in Southern India and proposed for it the name *laterite* (L. later=brick). The term has been widely applied to tropical red soils of widely differing characters and no problem in soil genesis has been more earnestly debated than the nature and mode of origin of this material. The study of tropical soils is, indeed, dominated by the laterite problem. Much of the confusion has arisen through inability to distinguish between soil as a material and soils as individuals. So long as it is thought possible to define a soil by the analysis of a sample, any soil showing a high proportion of sesquioxides may be considered a laterite. Indeed, F. J. Martin and H. C. Doyne proposed to define laterite as a material in which the molecular ratio of silica to alumina in the clay is less than 1.33, whilst in lateritic soils the ratio is 1.35 to 2.0. Apart from the fact that such a definition neglects iron as a constituent of the clay complex, it would have the result of bringing into one group soils of fundamentally different profile characteristics and previous history. The term thus used merely connotes a material. Confusion might be avoided in descriptions of tropical soils if the term *ferrallitic* were substituted for *lateritic* when the intention is merely to convey the information that the
material described is highly sesquioxidic. There are, nevertheless, good reasons for applying the term to certain types of profile occurring in the tropics.

The laterite controversy raises some philosophical problems that may be more appropriately discussed in a later chapter. Much of the confusion has arisen through the assumption, as the writer holds, mistaken, that there is an ideal or universal laterite and that the task of the student of tropical soils is to discover the characteristics or, in the language of the schoolmen, the "accidents" by which it may be recognized. Some have thought to recognize these accidents in the morphology of the profile, some in the chemical composition of the soil material, while some, the most uncritical, have even thought to recognize laterites in tropical soils with red colours. If the term laterite is to be used in pedology, it would probably be safest to adopt a frankly empirical position and propose that the term shall be restricted to one alone of the many types of soil actually occurring in the tropics.

On grounds of practical convenience, the ideas of R. L. Pendleton²⁷ may be commended. This writer would restrict the use of the term laterite to profiles characterized by the presence of concretionary material or crusts overlying mottled or vesicular horizons, and presumably developed under ground-water influence as first described by J. M. Campbell²⁸ in W. Australia and Africa. The application of the term would thus be denied to ferrallitic material of primary weathering such as the Nipe Clay of Cuba and similar soils developed from ferrallitic materials under conditions of free drainage. Pendleton follows Campbell in considering laterites to have been developed originally as pseudo-illuvial deposits in the zone of a fluctuating watertable under peneplastic conditions. On later uplift and dissection, the ground-water conditions disappeared and a new direction of profile development was induced. Profiles whose characteristics may be explained on this hypothesis occur in many parts of the world.
The description of a typical laterite profile as given by H. Harrassowitz exhibits some of the principal features of this soil group. According to this writer, the profile consists essentially of four horizons, namely, (1) the parent rock, succeeded by (2) an horizon representing the material of primary weathering, which appears to be of a kaolinitic character in turn succeeded by (3) the lateritic horizons, passing into (4) a surface horizon characterized by ferruginous incrustations or concretions. The zone of decomposition may be a mere film or may attain a thickness of a metre or more.

Harrassowitz's lateritic horizon consists of red material mottled with yellow or even violet, generally argillaceous, but when derived from quartzose rocks, often sandy in texture, it has a tendency to a vesicular or cellular structure and the pores are frequently filled with white or greyish material. When moist, it may be dug out, but hardens on drying so that it may be used as a building material. The surface ferruginous crust may attain such a development as to preclude the growth of vegetation. With it is associated free alumina as gibbsite or hydrargillite.

C. G. Stephens has studied pedogenesis following laterite formation in Southern Australia. Adopting Campbell's views as to their original formation he considers these laterites to have been developed during Tertiary times. More recently, uplift, dissection and new profile development have taken place. He describes normal and truncated profiles and considers the normal profile, developed from the original laterite, as essentially podsolic. The characteristic features surviving from the original laterite formation are (1) the horizon of nodular or pisolitic laterite gravel, underlain by (2) clay with mottled yellow, red and yellowish grey colours. The superficial bleached sandy horizons are presumably of later and roughly contemporary origin.

R. L. Pendleton (loc. cit.) has described laterites in Siam, where they are widely developed. He follows Campbell in considering them to be ground-water formations.
and distinguishes two processes, namely (1) those occurring in deep horizons where the geological parent rocks are permanently saturated with water; in this zone, iron compounds go into solution; and (2) those occurring above this zone, where ferrous compounds become oxidized through access of air. This oxidation takes place in the moist porespaces above the water-table giving the characteristic mottled appearance of the laterite horizon. R. L. Pendleton and S. Sharasuvana give numerous analyses of lateritic profiles from which it appears that the concretionary material, whilst generally sesquioxicidic in character, may contain considerable proportions of silica. The presence of superficial bleached horizons has been noted in many lateritic profiles, which some authors have described as podsolized. Pendleton considers that such bleaching differs entirely from podsolization, which occurs in the presence of raw humus, and is more properly termed lixiviation following the usage of E. C. Mohr.

Whilst the typical lateritic profile would appear to be marked by the presence of an horizon of concretionary or crust-like material underlain by mottled clay, there are also profiles in which, although concretionary material occurs, the mottled horizons are absent. Such a type is described by F. Hardy and R. R. Follett-Smith. It occurs in British Guiana and is derived from hornblende-schist. Their description is as follows:

1. Superficial laterite and ironstone, frequently vesicular and slaggy, scattered in blocks on the surface.
3. Tawny red earth, friable and very gravelly, especially in its upper part. The gravel consists of pieces of laterite, ironstone pellets, and quartz fragments. Thickness 10-20 feet.
4. Primary red laterite, hard and crusty. Thickness 3 inches over broken rock surface.

Gibbsite (Al₂O₃·3H₂O) occurs throughout the profile. It would appear that in this type of profile a ferralllitic material is the primary product of weathering. F. J. Martin and H. C. Doyne also have recorded the direct formation of similar material from norite in Sierra Leone.
The parent material appears to have considerable significance, for, according to Hardy and Follett-Smith, acidic rocks such as granite do not yield laterites. This may be connected with the formation of a kaolinitic type of weathering product from the potash and potash-soda felspars occurring in such rocks. It will be remembered that the non-sesquioxidic “ilepa” soil of S. Nigeria, described by Doyne and Watson (p. 404) is derived from acidic rocks.

Great differences of opinion exist as to the mechanism of laterite formation. It is considered by J. B. Harrison that two main processes may be distinguished, namely, (1) primary lateritic formation and (2) re-silicification of the lateritic product by deposition of silica from ascending solutions. The latter process only takes place in localities affected by a high water-table. Otherwise, the laterite appears to be permanent.

F. Hardy and R. R. Follett-Smith (loc. cit.) adduce evidence for a process similar to podsolization affecting the upper horizons of lateritic profiles, whereby sesquioxides are removed by leaching.

J. M. Campbell, from a study of Australian and African laterites, considered laterites to result from changes which occur in the zones of permanent and intermittent saturation in the vicinity of a water-table. Since oxygen is necessary for the oxidation of ferrous to ferric iron, the proximity of these zones to the surface is a further condition for its formation.

The hydrated sesquioxides deposited in the lateritic horizons originate by hydrolysis of silicate minerals and occur in the ground-water in alkaline solution, the iron as ferrous hydrogen carbonate and the aluminium as alkaline aluminates. During periods of drought, there is an upward translocation of dissolved material from the water-table. Ferrous hydrogen carbonate undergoes oxidation, giving amorphous hydrated ferric oxide, whilst the aluminates give rise to amorphous hydargillite $\text{Al}_2\text{O}_3 \cdot 3\text{H}_2\text{O}$.

Subsequent changes in the lateritic horizon involve the formation of turgite $2\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O}$ and possibly haematite,
Fe$_2$O$_3$, from amorphous ferric oxide hydrate, and crystalline gibbsite Al$_2$O$_3$.3H$_2$O from hydrargillite. On account of the greater mobility of ferric oxide in colloidal solution, it may happen that lateritic horizons become progressively more aluminous in character with age.

The views of Campbell are supported by the investigations of W. G. Woolnough$^{32}$ in Western Australia. This author considers that laterization can occur only in situations in which free drainage is at a standstill, as in a peneplain at or about sea level. High level laterites are products of processes which have long since ceased consequent on uplift and the alteration of hydrological conditions. Laterite, as imagined by these authors, was formed by a process which may be paralleled by the formation of "murrum" and "mocarrero" in contemporary vlei soils. Lateritic soils may, indeed, be considered as the tropical analogues of the gleys soils of temperate regions.

The formation of a material such as the Nipe clay, which appears to be mainly residual in character, must obviously be distinguished from the processes whereby profiles characterized by slaggy ferruginous incrustations are developed. In these, the more typical lateritic profiles, it is clear that there has been a deposition which has masked the primary processes of silicate hydrolysis and clay formation. It cannot be assumed in all cases, however, that profile development has been influenced by the proximity of ground-water, for contemporary laterites may occur under savannah. An alternation of extreme wetness followed by desiccation may have the same effect as a fluctuating water-table.

Whether or not illuvial deposition of sesquioxidic material has taken place in the presence or absence of a high ground-water table, it is evident that the principal climatic circumstances are high temperature and the alternation of aerobic and anaerobic conditions within the soil profile.

The mobility of sesquioxidic constituents in the soil profile has been attributed by Campbell to the alkalinity of the ground-water, by H. Harrassowitz$^8$ to the protective effect
of humus sols and by A. Reifenberg to the protective effect of silicic acid sols. It is probable that each of the three explanations may find an application in the wide range of lateritic profiles encountered.

VOLCANIC SOILS

This somewhat misleading name is given to soils derived from volcanic ashes and tuffs under tropical and sub-tropical conditions. The nature of the parent material is such that it weathers very rapidly. F. Hardy reports that andesitic ashes, accumulated during the eruptions at St. Vincent, B.W.I., in 1902-3, were able to carry agricultural crops within one year with the aid of small organic dressings. Similar instances of the rapidity with which such materials become available for agricultural use can be found in many parts of the world. The character of the ash is of importance. Whilst highly acid ashes weather slowly, basic ashes weather rapidly to fertile soils. N. H. Taylor reports volcanic ash soils from New Zealand of an aluminous character.

GENERAL CONSIDERATIONS ON TROPICAL SOILS

Some of the general characteristics of tropical soils were mentioned at the beginning of this chapter. Many of these distinctive features might be regarded as geological rather than as pedological. The occurrence of highly weathered parent materials, often markedly ferrallitic in character, and the extensive erosion and re-sorting of the superficial deposits that has taken place are all distinct from the processes of pedogenesis, if by this term is understood the development of the profile. On the other hand the characteristic but little understood organic matter regime, the eluvial processes, and the tendency to concretionary developments in the profile are essentially connected with pedogenesis.

In spite of these marked peculiarities of tropical soil formation the affinities of tropical soils with those of temperate regions must not be overlooked. The tropical red loams can be regarded as the analogues of the temperate brown earths, the tropical laterites of the temperate gley
soils, and the bleached surface layers sometimes encountered betoken a process akin to podsolization.

The problems that confront the pedologist in the tropics are as much geological as pedological. How closely geology and pedology are associated in producing the tropical soil pattern may be readily seen from a perusal of the late Geoffrey Milne’s *Soil Reconnaissance Journey through Parts of Tanganyika Territory*. Unfortunately, the geologist is mainly concerned with the solid geology, and the necessary interpretation of regolithic phenomena must be the responsibility of the pedologist.

**BIBLIOGRAPHY**


SOILS OF THE HUMID TROPICS


CHAPTER XV

SOILS ASSOCIATED WITH CALCAREOUS PARENT MATERIALS

The soils to be discussed in the present chapter are grouped together principally for descriptive purposes and not, necessarily, because they form a major group in a world system. Their position in such a system will be discussed at a later stage.

Soils derived from or associated with calcareous parent materials, such as limestone, chalk, or coral, appear to fall into two main groups. On the one hand, is a group of soils in which the mineral colour is grey or greyish-brown, and the colour of the soil as a whole tends towards brownish or greyish-black shades. With them may be grouped white or grey chalk soils in which the organic matter colour is masked by the excess of chalk. On the other hand is a group of soils in which the colour is predominantly red or reddish-brown, sometimes modified by the presence of organic matter, which would appear to be of a different character from the organic matter of the soils of the first group. The clay complex of the black and grey soils is more siliceous than that of the red and brown soils, which appears to contain free sesquioxides. Soils of the first group contain free calcium carbonate and are completely saturated with exchangeable bases, whilst soils of the second group generally show a distinctly lower base-status and may be actually acid in reaction. To the first group belong the rendzinas, with which may be included the grey and white limestone soils of the temperate regions, and the black limestone and marl soils of the tropics. To the second group
belong the so-called *terra rossa* soils, with which may be included the brown and red limestone soils of temperate regions and the red limestone soils of the tropics.

**RENDZINAS AND RELATED SOILS**

These soils have been most widely studied in Central Europe. They are sometimes described as humus-carbonate soils. They are, typically, dark coloured soils containing from 3-12% of organic matter and varying amounts of free calcium carbonate. In Glinka’s system of classification, they are considered to be typical endodynamomorphic soils, which have not yet reached mature development and in which the profile character is mainly determined by the nature of the parent material. Though generally of a dark colour, soils belonging essentially to this class may also be dull brown or even grey in colour. Reddish-brown colours do not occur.

W. Kubiena distinguishes “proto-rendzinas” and “mull rendzinas,” the former being of a skeletal type without earthworms and the latter having well developed mull with abundant earthworms.

According to K. D. Glinka the essential features of a rendzina profile are:—

A. Grey, dark grey, or almost black soil, sometimes containing fragments of chalk or marl. Thickness 15-30 cm.

A. Whitish-grey, slightly humus coloured or brownish, with more fragmental chalk or marl than in the A, horizon.

C. Calcareous parent rock.

L. G. Kotzmann gives the following as the typical features of a Central European rendzina profile developed on pure limestone or dolomite:—

A. Dark grey or almost black soil with loose crumbly structure. Different-sized fragments of limestone or dolomite, sometimes coloured with iron, soil generally sharply delimited from the parent rock, free of humus and contamination. Thickness varies from 20 to 30 cms. Organic matter content up to 20-25%.

C. Upper part of parent rock generally consists of pink-coloured fragments increasing in size with depth (C1). Below it is unaltered limestone or dolomite.
The clay fraction is generally carbonate-free and appears to consist mainly of quartz and the weathering complex, which is of a siliceous character.

*Poland.* S. Miklaszewski distinguishes two groups of rendzinas, namely, those derived from calcareous parent materials and those from gypseous parent materials. Among the calcareous rendzinas, white, brown, and black varieties occur. The colour is not always correlated with humus content, for an instance is given of a white rendzina, containing 3.5% of humus, adjacent to a black rendzina with 2.8% of humus. The white rendzinas are probably similar to the white chalk soils of England, in which the organic matter colour is masked by the excess of chalk. The presence of considerable proportions of calcium carbonate is a constant feature of typical rendzinas.

*England.* There is now fairly general agreement in assigning the chalk soils of England to the rendzina group. The prevalence of grey or brownish-grey colours and the presence of free calcium carbonate would appear to warrant their inclusion. Even in the light-coloured chalk soils, the residue after removal of calcium carbonate has the dull greyish-black colour characteristic of organic matter formed in the presence of excess of calcium carbonate.

The chalk soils of Great Britain generally appear to have undergone considerable erosion, and this has tended to obscure their relationship to the continental rendzinas. Yet, under certain conditions, profiles occur in which there is a well marked greyish-brown or even black surface horizon.

A chalk soil profile, occurring under arable cultivation near Wantage, Berkshire, is described by Miss F. F. Kay in the following terms:

- **0.10 in.** Grey fine sandy silt clay loam; very calcareous, loose.
- **10-16 in.** Putty-coloured fine sandy silt loam containing many fragments of chalk rock, which increase in size with depth.
- **>16 in.** Chalk (Lower Chalk) rock.
The following analytical data were obtained on samples of the above profile (Table XLII).

**Table XLII.—Analytical data for Wantage soil profile**

<table>
<thead>
<tr>
<th>Horizon</th>
<th>CaCO₃ %</th>
<th>Organic Carbon %</th>
<th>Clay Fraction</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>SiO₂/Al₂O₃</td>
<td>SiO₂/R₂O₅</td>
</tr>
<tr>
<td>0-10 in.</td>
<td>21.0</td>
<td>2.2</td>
<td>5.65</td>
<td>4.47</td>
</tr>
<tr>
<td>10-16 in.</td>
<td>36.1</td>
<td>0.65</td>
<td>5.95</td>
<td>4.63</td>
</tr>
</tbody>
</table>

The same writer gives the following description of a dark-coloured rendzina under old pasture on the downs near Newbury:

0-5 in. Dark brown organic friable loam (turf layer); crumb structure, loose.

5-7 in. Layer of medium sized sub-angular flints.

>7 in. Brown stained chalk fragments and flints passing into unbroken chalk (Upper Chalk) rock.

The following analytical data were obtained for the top layer:

<table>
<thead>
<tr>
<th>CaCO₃ %</th>
<th>Organic Carbon</th>
<th>Clay Fraction</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>1.45</td>
<td>9.6</td>
<td>SiO₂/Al₂O₃ 3.16</td>
<td>SiO₂/R₂O₅ 2.47</td>
</tr>
</tbody>
</table>

The clay fraction is less siliceous than in the Wantage soil, probably reflecting the differing character of the parent chalk.

**Wales.** Certain soils derived from the Lower Lias would appear to belong to the rendzina group. The following is a description by D. O. Hughes⁶ of a typical profile in Glamorganshire under old pasture:

0 to 6 in. Greyish-brown stoneless clay loam.

6 to 15 in. Light yellowish-brown clay, with occasional fragments of limestone.

>15 in. Soft limestone.
The following analytical data were obtained (Table XLIII).

**Table XLIII.—Analytical Data for Lias Soil Profile**

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Clay %</th>
<th>CaCO₂ %</th>
<th>Organic Carbon %</th>
<th>Clay Fraction</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>SiO₂/Al₂O₃</td>
</tr>
<tr>
<td>0-6 in.</td>
<td>39.5</td>
<td>2.7</td>
<td>3.3</td>
<td>4.50</td>
</tr>
<tr>
<td>6-15 in.</td>
<td>44.0</td>
<td>17.7</td>
<td>1.2</td>
<td>4.29</td>
</tr>
</tbody>
</table>

The highly siliceous character of the clay fraction is a distinctive feature of soils belonging to this group which have been examined by the writer. On theoretical grounds, the same character may be expected in all soils containing an excess of calcium carbonate, maintaining a state of base-saturation.

The writer has encountered a dark-coloured pasture soil over Carboniferous Limestone near Penmon, Anglesey, consisting of 5 inches of black crumbly humous loam with abundant limestone fragments over shattered limestone.

**Cuba.** The Camaguey series of Cuba described by H. H. Bennett and R. V. Allison⁷ may serve as an example of a tropical black limestone soil. Typically it consists of (1) 6-14 in. of dark brown to black clay, overlying (2) dark brown to yellowish-brown plastic clay, frequently calcareous, grading at 20-30 in. into (3) brownish-yellow to yellow, plastic and somewhat sticky clay, frequently calcareous and containing lumps of soft calcium carbonate, overlying at 40-72 in. (4) either whitish cocó (chalky limestone), pale yellow calcareous clay and white cocó, or cocó and calcareous clay, interbedded with hard or semi-hard limestone.

The analytical figures given show that calcium carbonate is present throughout the profile. The organic matter in the surface soil is 9.62%.

**Barbados.** S. J. Saint⁸ describes black soils derived
from coral limestone, which appear to be of the rendzina type. The following is typical.

0-12 in. Black surface soil.
12-24 in. Black subsoil with greyish-white streaks.
24-36 in. Greyish-white soapy clay; black streaks and concretions.

The analytical figures are as follows:

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Mg%</th>
<th>CaCO₃</th>
<th>Clay</th>
<th>SiO₂/F₃O₅</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-12 in.</td>
<td>1.69</td>
<td>4.8</td>
<td>60.7</td>
<td>3.34</td>
</tr>
<tr>
<td>12-24 in.</td>
<td>0.67</td>
<td>3.7</td>
<td>68.6</td>
<td>3.36</td>
</tr>
<tr>
<td>24-36 in.</td>
<td>0.58</td>
<td>5.8</td>
<td>73.0</td>
<td>3.14</td>
</tr>
</tbody>
</table>

MAGNESIUM RENDZINAS

V. Novák and J. Pelišek describe soils derived from serpentine in Moravia which appear to be related to the rendzinas. They are alkaline to the surface and have a very stable structure. The parent rock contains only a trace of calcium, but has 38.6% MgO. This is reflected in the character of the exchange complex in which about 80% of the exchangeable cations are represented by magnesium. They are thus definitely magnesium soils in contrast to ordinary rendzinas, which are calcium soils. Novák and Pelišek term them magnesium rendzinas.

RED AND BROWN LIMESTONE SOILS

Terra rossa is the name given to a red soil which occurs commonly in the countries bordering on the Mediterranean Sea. Typically it is associated with limestone and in the present account we shall follow the definition of A. Reifenberg in restricting the term terra rossa to such soils.

The soils included in this group show a simple type of profile consisting of a variable depth of red soil sometimes slightly modified in colour by admixture of humus in the upper horizon, and passing abruptly to the limestone rock below without any marked zone of transition. Typically, the terra rossa soils are fairly heavy clays. Analyses by W. Graf zu Leiningen show clay contents varying from
32.85 to 59.05% in typical examples. The coarser fractions are correspondingly small. In some cases, fragments of the parent rock and also calcareous concretions stained red with iron oxides are present.

Although numerous bulk analyses are found in the literature, scarcely any data are available for the composition of the clay fraction. A clay from Pikermi examined by zu Leiningen gave a clay fraction having the composition,

\[
\begin{align*}
\text{SiO}_2 & \quad 41.62 \\
\text{Al}_2\text{O}_3 & \quad 21.16 \\
\text{Fe}_2\text{O}_3 & \quad 12.85 \\
\text{CaO} & \quad 3.82 \\
\text{Ignition} & \quad 15.05
\end{align*}
\]

Similar figures were obtained by E. Blanck, F. Kunz, and F. Preiss\textsuperscript{12} for certain red soils from Moravia with resemblances to terra rossa. The present writer has made partial analyses of the clay fractions of two red soils over limestone from Cap d’Antibes, S. France, and St. Gowans Head, Pembrokeshire, respectively. The figures are as follows:

<table>
<thead>
<tr>
<th>Clay Fraction</th>
<th>SiO\textsubscript{2}/Al\textsubscript{2}O\textsubscript{3}</th>
<th>SiO\textsubscript{2}/R\textsubscript{2}O\textsubscript{3}</th>
<th>Al\textsubscript{2}O\textsubscript{3}/Fe\textsubscript{2}O\textsubscript{3}</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cap d’Antibes</td>
<td>2.61</td>
<td>2.04</td>
<td>3.19</td>
</tr>
<tr>
<td>St. Gowans Head</td>
<td>2.51</td>
<td>1.95</td>
<td>3.51</td>
</tr>
</tbody>
</table>

Neither the abundant data for the bulk composition of terra rossa soils nor the few available data for the composition of the clay fraction suggest any tendency towards ferrallitization. The figures are such as might be obtained from typical brown earths, or younger red tropical loams.

Generally speaking, terra rossa soils are base-un-saturated. Calcium carbonate is, indeed, shown in certain analyses, but it is not improbable that it may be present in the form of fragments which, by reason of their size or degree of induration, are without effect on the base-status of the body of the soil. In this connexion, it has been remarked by K. D. Glinka, that whilst soft limestones produce rendzina soils, hard limestones produce red soils. If the development of red soils be a consequence of a lowering of base-status, it is readily seen that such a process can most readily take place in soils derived from hard limestones.

The problem of the origin of terra rossa has been
earnestly debated both by pedologists and geologists. The theory that it is simply a residual product from the parent limestone is inadequate, since material resembling terra rossa cannot be obtained by dissolving away the carbonates from limestones. And, therefore, whilst its close association with the limestone suggests an origin from the non-carbonate portion of the rock, it is evidently necessary to account for the changes whereby this material has become terra rossa.

H. Stremme\textsuperscript{13} considers terra rossa to be analogous to the B horizon of podsol profiles, the A horizon having been removed by erosion. But whilst red subsoils similar to terra rossa occur in certain parts of Central Europe under humus surface soils, it does not appear possible to apply this theory of their origin in the Mediterranean region generally.

A. Reifenberg\textsuperscript{7}, from an intensive study of the conditions of terra rossa formation, concludes that the essential feature in its formation is the irreversible precipitation, in the surface horizons, of sesquioxide soils peptized by silicic acid. Such data as are available, however, do not suggest a high content of sesquioxides for terra rossa, and it may be more correct to regard these soils as analogous to the brown earths, in which, if we accept the views of A. Stebbutt, the lowering of the base-status has resulted in a partial degradation of the zeolitic or clay complex into its component silicic acid and sesquioxides.

A full discussion of the problem of terra rossa is given in Blanck's *Handbuch der Bodenlehre*, Vol. III. (pp. 193-257). A more recent discussion is given by W. Hollstein\textsuperscript{14}. Perhaps the most baffling problem is that of the origin of the iron oxide. It has been remarked that the best developed terra rossa soils are on the purest limestones, and it may well be that the iron is of external origin, possibly from volcanic dust. A relationship with the red clay of the greater oceanic depths would be worth investigation.

*Cuba*. As an example of a tropical red limestone soil, the Matanzas series of Cuba described by H. H. Bennett and R. V. Allison\textsuperscript{15} may be cited.
The profile is comparatively simple and consists of a uniform red friable clay, which may reach to 15 feet or more with an abrupt transition to the parent rock. The top 3-6 inches is somewhat darker in colour through the presence of organic matter. A characteristic of this series is the occurrence of concretionary material (perdigón). The parent rock is light coloured limestone, locally flinty, or limestone and cocó.

The analytical figures presented show a complete leaching of calcium carbonate, a condition which is reflected in the acid reaction of the soils of this series. The organic matter content in the three surface soils for which data are given varies from 2.19 to 4.65%. Although no data for the composition of the clay fraction are available, free sesquioxides are probably present. The concretions consist mainly of ferric oxide with smaller proportions of alumina and silica.

United States. The Hagerstown series which is found in the Blue Ridge region of the United States belongs to the group of red and brown limestone soils. The soil profile, where mature, consists of a light-textured light brown A horizon and a brown or reddish-brown B horizon. Calcium carbonate has been entirely removed from the profile by leaching. Another example of this group is the Decatur series of Georgia and neighbouring states.

Wales. Brown, red, and reddish-brown soils are commonly found on hard Carboniferous Limestone in Wales under conditions which have resulted in decalcification and desaturation. It is noteworthy that, under the same climate, the soft Lias Limestone gives rise to greyish-brown soils. The following profile is found on Carboniferous Limestone at Bridgend, Glamorgan:

0-9 in. Rusty reddish-brown silty heavy loam.
9-15 in. Brown silty heavy loam with fragments of limestone.
>15 in. Hard limestone.

Fragments of limestone occur in both horizons and markedly in the second horizon. Yet the soil is partly leached, for the pH is 5.96 and the exchangeable lime
figures, 0.379% and 0.313% respectively for the two horizons, indicate partial desaturation. The composition of the clay fractions is as follows:

<table>
<thead>
<tr>
<th></th>
<th>SiO₂/Al₂O₃</th>
<th>SiO₂/R₂O₅</th>
<th>Al₂O₃/Fe₂O₃</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-9 in.</td>
<td>2.50</td>
<td>2.07</td>
<td>4.81</td>
</tr>
<tr>
<td>9-15 in.</td>
<td>2.76</td>
<td>2.17</td>
<td>3.07</td>
</tr>
</tbody>
</table>

These figures show a marked contrast with those for neighbouring greyish-brown Lias soils, which appear to belong to the rendzina group.

Whilst in ordinary Welsh profiles, the tendency is for sesquioxides to increase in the clay fraction from the surface downwards, there is here no evidence of eluviation in either direction. The Al₂O₃/Fe₂O₃ ratio decreases with depth.

The fairly high proportions of organic matter in the desaturated limestone soils of Wales tend to mask their mineral colour, which is often decidedly brick-red. More commonly the colour is a warm reddish-brown.

**RELATIONSHIP OF RENDZINAS TO RED AND BROWN LIMESTONE SOILS**

We have seen that limestones can give rise to two distinct groups of soils. The one group is characterized by a high base-status, excess of calcium carbonate, a dark coloured type of organic matter, and a grey or greyish-brown mineral colour associated with a siliceous type of clay, probably devoid of free sesquioxides.

The other group is characterized by a generally low base-status, sometimes even in the presence of fragmentary limestone, a type of organic matter of a lighter colour than that of the first group, and a red or reddish-brown mineral colour, associated with a type of clay having a SiO₂/R₂O₅ ratio of about 2.0 or less, in which there is a certain proportion of free sesquioxides.

The formation of the second group appears to follow the desaturation of the soil, and we may therefore regard the red limestone soils as more mature than the grey or rendzina.
soils. In a limestone country, it should be possible to dis-
cern different stages of maturity due to variations in topo-
graphy. If mature leached soils are to be found, they may
be expected in upland situations with relatively high rainfall
and intense leaching, provided erosion is not severe enough
to keep pace with profile development.

An example of this is furnished by red and black soils
derived from coral limestone in Barbados, described by S. J.
Saint. The red soils occur in the districts of highest alti-
tude and greatest rainfall, whilst the black soils occur in the
drier lowlands. All intermediate gradations are found
between the extremes. The black soils are more colloidal
than the red soils, and have a more siliceous clay fraction.
Thus, the clay $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio for the upland red soils is
1.95, whilst the corresponding figure for the lowland black
soils is 3.29. It is rather remarkable that the red soils
have a higher organic matter content than the black soils.
The dark colour of the latter soils may be associated with the
higher calcium status of the humus, a circumstance which
has been remarked elsewhere; but Saint considers it to be
of mineral origin. According to Saint, the key to the dis-
tinction between the two types lies in the greater maturity
of development of the red soils.

J. M. Albareda and E. Gutiérrez Ríos, from a study of
soils derived from limestone in Spain, conclude that the
Spanish red soils, which may be regarded as belonging to
the terra rossa class, have been formed by the degradation
of rendzinas consequent on deforestation under semi-arid
conditions. Under forest the tendency of evolution would be
towards brown earth or podsolic types.

In the writer's experience, soils derived from hard lime-
stones in Wales under conditions of intense leaching always
show reddish or brownish colours, whilst colluvial soils or
soils in situations with imperfect drainage are generally
greyish or brownish-grey. Some figures recently obtained
for an Anglesey profile in limestone drift support the view
that desilicification accompanies decalcification. The soil
down to 30 inches is devoid of calcium carbonate and the clay 
$\text{SiO}_2/\text{R}_2\text{O}_3$ ratio is 1.85 for each 15-inch layer. Below 30 
inches the material is calcareous and has a clay $\text{SiO}_2/\text{R}_2\text{O}_3$ 
ratio of 3.90.

In view of these considerations, it may be permissible to 
hazard a conjecture that the red Clay-with-flints overlying 
certain parts of the Chalk of S.E. England may represent a 
mature limestone soil. Against the view that it is a residual 
product from chalk it has been urged that the chalk 
which has been weathered away could not have produced a 
layer of Clay-with-flints of the thickness which actually 
occurs. Yet a considerable proportion of the clay must be 
chalk residual material, and it is not impossible that some 
additions may have been made by illuvial accumulation as 
has been suggested for the terra rossa.

It may be remarked that mature limestone soils tend to 
resemble the soils of warmer climates. The Mediterranean 
terra rossa shows points of resemblance with the red loams 
of the tropics, whilst the red colours of leached limestone 
soils of more northern climates are in marked contrast to the 
more sober brown colour of adjacent soils derived from non-
calcareous parent materials. This may be attributed to the 
drier and, therefore, warmer character of these soils and to 
the more rapid mineralization of plant residues.

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CHAPTER XVI
THE CLASSIFICATION OF SOILS

INTRODUCTORY

Attempts to classify soils have been made from the earliest days, and in this matter the practical man has preceded the scientist. Farmers from the earliest times have recognized different kinds of soil through hard experience and have given them names usually expressing their differing textures. In modern times the connexion between soil and parent material was soon recognized, and one of the earliest attempts to find a scientific basis for classification was that of F. A. Fallou, who, from his experience in Saxony, proposed a petrographical-geological system. This and most of the earlier essays in classification suffered from their limited geographical scope and from their inadequate knowledge of the different factors, such as climate and topography, operative in soil development. The significance of the soil profile has been generally recognized only within comparatively recent years.

Before entering on a detailed consideration of the different systems of soil classification, it may be helpful to examine some of the philosophical problems involved, in order to understand the general limitations inherent in all attempts to group soils into ordered systems.

If we wish to describe the soils of a farm, of a parish, of a country, or indeed of the whole world, some kind of a classification or ordered arrangement is necessary. In classifying soils, as in classifying any other objects, many distinct methods are conceivable. For example, we might classify soils according to their colour, their organic matter content, their texture, and so on, and each of these classifi-
cations might have a certain value. We may distinguish two contrasted approaches to the problem. On the one hand, we may base our classification of soils on the various factors that have contributed to their formation, such as parent material, climate, and topography; on the other hand we may make our classifications in terms of observable soil properties. The first type of classification is exemplified by petrographical, climatic, and vegetational classifications, the second by classifications based on profile characteristics. If we use the term "property" in its widest sense to include not only observable characters but also the factors known or supposed to have been operative in forming the soil, we might define a class X in a particular system as being a number of soils having properties A, B, C, D, etc., in common. A classification giving classes, in which all members of each class have all properties in common, is impracticable with such a variable object as the soil. And therefore, within a class of soils, there must always be a number of properties in which differences are shown. The problem consists in the selection, perhaps not exhaustive, of properties common possession of which shall define the class.

Here it is necessary to distinguish properties according to their significance and in this connexion it is relevant to quote some remarks of J. S. Mill (System of Logic, 1879, Vol. II., pp. 270, 272) on the subject of classification:

"Classification, thus regarded, is a contrivance for the best possible ordering of the ideas of objects in our minds; for causing the ideas to accompany or succeed one another in such a way as shall give us the greatest command over our knowledge already acquired, and lead most directly to the acquisition of more."

"The ends of scientific classification are best answered, when the objects are formed into groups respecting which a greater number of general propositions can be made, and those propositions more important, than could be made respecting any other groups into which the same things could be distributed. The properties, therefore, according to which objects are classified, should, if possible, be those which are causes of many other properties, or, at any rate, which are sure marks of them. Causes are preferable, both as being the surest and most direct of marks, and as being themselves the properties on which it is of most use that our attention should be strongly fixed. But the property which is the cause of the chief peculiarities of a class is unfortunately seldom fitted to serve also as the diagnostic of the class.
Instead of the cause, we must generally select some of its more prominent effects, which may serve as marks of the other effects and of the cause."

The selection of the properties for use as criteria in classification is not always easy, for their significance may not be plainly apparent. Thus, mechanical composition may appear very significant because of its importance in practice, yet the grouping of soils into classes with similar or identical mechanical composition would bring together soils that would differ from each other in most important respects. The aims of a practical classification and those of a scientific classification may run counter to each other. Thus, for certain practical purposes we may group together soils that are widely different from the scientific standpoint, just as the term "roots" in crop husbandry includes crops belonging to the Cruciferae, Chenopodiaceae, and Solanaceae—widely separated natural orders.

Contrasted with classifications based on the observable properties of soils, there are those based on external circumstances or factors that have contributed to soil formation. Many classifications based on soil-forming factors have been proposed and all have a certain value. Strictly speaking, these are not classifications of soils but of the factors concerned in the formation of soils. H. Jenny (Factors of Soil Formation), indeed, prefers to express the known variations in soils by a series of equations expressing the effect of different factors on soil properties. Such an approach to the problem would dispense with the recognition of definite classes and arrange soils in sequences. (Cf. p. 458.)

Jenny criticizes the whole idea of grouping soils into classes and prefers to characterize soils by functional relationships based on equations showing individual soil properties as functions of the different pedogenic factors. Speaking of soil maps he remarks ("Factors of Soil Formation," p. 262):

"The goal of the soil geographer is the assemblage of soil knowledge in the form of a map. In contrast, the goal of the "func-
tionalist’ is the assemblage of soil knowledge in the form of a curve or an equation. These objectives are clearly brought out in a comparison between Marbut’s *Soils of the United States* and the present treatise on soil-forming factors. The former abounds with carefully executed maps of various scales, whereas the latter displays a sequence of graphs and equations. Both methods of approach have their merits, and nothing would be gained by playing one system against the other. Soil maps display the areal arrangement of soil properties and types but give no insight into ‘causal’ relationships. The curve, on the other hand, reveals the dependency of soil properties on soil-forming factors, but the conversion of such fundamental knowledge to specific field conditions is impossible unless the areal distribution of the soil formers is known. Clearly, it is the union of the geographic and the functional method that provides the most effective means of pedological research.”

If we accept these ideas, the boundaries shown on a soil map are largely arbitrary and merely mark the occurrence of conventional points on a series of continuous curves. Yet we cannot entirely abandon the idea of soil boundaries or the possibility of representing soils cartographically. Whilst successions such as chernozem to desert or podsol to brown earth may be so gradual as to preclude definite boundaries, in other cases definite boundaries plainly exist. A change in parent material often results in a perfectly definite and mappable boundary, as may be seen over many parts of Britain. Similarly, topography may result in a sharp transition from soils of free drainage to soils of impeded drainage. The practical soil surveyor, although aware of the occurrence of transitional soils and merging boundaries, cannot but be aware of the existence of definite and mappable soil classes. It is, however, necessary to regard soil maps critically and to distinguish between soil boundaries of objective significance and those based on a subjective interpretation of the field data. We should also recognize another way in which soil maps fail to correspond exactly with the physical facts, namely, where generalized boundaries have been drawn in regions of great cartographic complexity. A soil map at best is only an approximate picture.

The problems of soil classification illustrate and are illustrated by the important philosophical problem of
*universal* and a short digression on this problem may be helpful to the student of soil.

If we consider any common object such as chair or table, it is clear that it is known to us only through sense-perceptions such as hardness, colour, shape, and so on. This led Plato to his theory of ideal forms. A particular table is merely a manifestation to our sense of an ideal table. We need not follow Plato into his world of ideals, which has in it, perhaps, as much of poetry as of philosophy. The Platonic ideas formed the basis of early medieval scholasticism which, following St. Augustine, accepted the belief in ideal forms or, as we should now say, universals, existing independently of their material manifestations. Whether it was possible to know these forms directly was a matter for controversy. This extreme position known as Augustinian Realism was modified as the ideas of Aristotle permeated western thought, at first through Boethius and later through the Arabian philosophers. In its later form as developed by St. Thomas Aquinas, universals were considered as existing not antecedent to, but in the objects to which they referred.†

Against Realism was the opposite doctrine of Nominalism, according to which universals are merely names and not independent facts. The issue was raised by Boethius in a commentary on Porphyry when he enquires whether genera and species exist in the same way as objects perceived by the senses. Are they not rather the products of our own thought and without counterpart in the real world? Intermediate between Nominalism and Aristotelian-Thomist Realism is Conceptualism which, whilst denying universals, admits the possibility of concepts, framed arbitrarily by the mind by abstraction from particular impressions.

*The following works have been found helpful in furnishing the ideas underlying this discussion: Bertrand Russell, Problems of Philosophy and History of Western Philosophy; M. H. Carré, Realists and Nominalists; J. Maritain, Introduction to Philosophy. In preparing this discussion I have been helped by discussion with my friend Professor R. I. Aaron, who does not necessarily subscribe to the views expressed therein.*

†According to St. Thomas, the abstraction of the universal (species intelligibilis) from the concrete particulars is the result of the operation of the active understanding (intellectus agens).
These old controversies become alive if we apply them to our thoughts about classes or kinds of soil. Has the "idea" of podsol an existence independent of our minds or does it merely mean a name applied to a range of soil profiles which, from our limited acquaintance with them, we believe to have enough in common to warrant the application of a common name?

The philosophical reader may recall the discussion in Plato's *Republic* on the meaning of justice. In this discussion there appears to be an underlying assumption that there exists an abstract ideal of justice and that this ideal is recognizable as embodied in its particular manifestations. That Socrates is able to confound Thrasymachus may seem to suggest that the conception of justice does exist if not as an innate ideal at least as an ideal that can be elicited and recognized. So in the controversies on the definition of such terms as laterite and podsol there appears to be an implied assumption that there is an ideal or universal laterite and an ideal or universal podsol and that these can be elicited and defined from experience.

When the student of soils becomes first acquainted with the term podsol, he tends to assume unconsciously a realist position and suppose that all actual podsols are particular manifestations of an ideal podsol. Hence we have the controversies centred round the definition of the podsol; for example, whether this or that character is critical in defining a podsol.

Such a belief may be untenable, because each student's conception of a podsol has been built up from his own direct or indirect experience. The podsol of Pedologist X differs from the podsol of Pedologist Y. Indeed, the same may be said of many of our conceptions, including some of the most familiar, such as table, house, plough, and motor car. We cannot, therefore, it would appear, define our ideal podsol simply from an analysis of experience or assume, with some,
that a definition is unnecessary by appealing to current usage.*

If there are certain common qualities, possession of which entitles us to speak of a class as podsols, we must consider the means by which we become aware of this. In actual experience, every student's conception of a podsol undergoes development. In the case of most British pedologists, first acquaintance is through description. In the second stage, soils are encountered in the field whose characteristics appear to agree with those learnt by description. The two most salient characteristics are, perhaps, the presence of a bleached A horizon and of a more or less ferruginous (and, by inference, sesquioxidic) B horizon. Where the student meets these characteristics he assumes that he is seeing a podsol. In the next stage, profiles possessing these characteristics are compared with a view to ascertaining what are the essential marks of a podsol. It may be found that some other characteristics are invariably present. Still other characteristics, e.g., the presence of a humic B horizon, may or may not occur. With wider experience, the student acquires a conception of a podsol that goes far beyond his original conception of a soil with a bleached A and a sesquioxidic B horizon.

If there are in a considerable group of soils a number of

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* Cf. F. P. G. Guizot, *Histoire de la Civilization en Europe* (Didier, 1859), p. 11: "... Il y a presque toujours, dans l'acception usuelle des termes les plus généraux, plus de vérité que dans les définitions en apparence plus précises et plus rigoureuses de la science. C'est le bon sens qui donne aux mots leur signification commune, et le bon sens est le génie de l'humanité. La signification commune d'un mot se forme successivement et en présence des faits; à mesure qu'un fait se présente, qui paraît rentrer dans le sens d'un terme comm., on l'y reçoit, pour ainsi dire, naturellement; le sens du terme s'étend, s'élargit, et peu à peu les divers faits, les diverses idées que, en vertu de la nature des choses mêmes, les hommes doivent rallier sous ce mot s'y rallient en effet. Lorsque le sens d'un mot, au contraire, est déterminé par la science, cette détermination, ouvrage d'un seul ou d'un petit nombre d'individus, a lieu sous l'empire de quelque fait particulier qui a frappé leur esprit. Ainsi, les définitions scientifiques sont, en général, beaucoup plus étroites, et, par cela seul, beaucoup moins vraies au fond que le sens populaire des termes." One should not conclude, however, that the popular idea of a podsol is more valid than a scientific definition.
clearly recognizable and clearly definable properties in common, then it would appear that there is a universal podsol, not in the Augustinian but in the Aristotelian sense, inherent in the group of soils and perceived and abstracted by the intellect. But here two difficulties occur: firstly there is the difficulty in the selection and perception of the critical properties, and secondly there is the difficulty of specifying these properties. In so far as the marks of a podsol do not permit of unambiguous definition the term podsol fails to represent a universal either in the Platonic-Augustinian sense or in the Aristotelian-Thomist sense, and may become no more than a mere name for a group of soils showing a general similarity. Yet the general notion of a podsol persists in the mind of the pedologist. It is framed as an abstraction from his impressions. Such an abstraction is largely arbitrary, i.e., the pedologist may select what he considers to be the most convenient criteria, subject to the limitation that these criteria are such as can be observed and defined. Precision is given by definition. This “conceptualist” standpoint is essentially practical and, if adopted, implies the drawing up of definitions for the classes of soils distinguished.

From what has been said, it would appear to be safest to regard soil classification rather as a convenient apparatus for grouping a highly complex and variable range of natural occurrences rather than as a given system of orders, genera, and species.* We should above all guard against assuming uncritically that the classes or categories such as

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*We must notice a fundamental difference between the genera and species of soils and the genera and species of plants. Even if we concede individuality to a soil, in that it is composed of an ordered arrangement of horizons forming a soil profile, it is entirely the result of certain pedogenic processes acting on a certain parent material. Although a plant or an animal may be profoundly affected by its environment, on which, indeed, its very existence depends, it cannot be regarded as entirely passive. Having inherited a certain chromosome pattern, it is, within limits, an active agent, with its own reaction to its environment. When we classify soils, we are really classifying the results of the action of a variety of environmental factors, including man himself and the lapse of time, on a variety of parent materials.
podsols, chernozems, etc., represent anything more than convenient labels for groups of soils possessing common and definable characteristics. From this, it would follow that each class should have its agreed definition, just as the terms used in an Act of Parliament are defined in the accompanying schedules. Until such definitions are generally agreed, whoever uses the term podsol without making clear the sense in which he is using the word is either falling into the error of the early medieval Realist or is using his terms loosely and without definition. Few pedologists can escape this accusation, and the present writer is no exception.

EARLIER GENETIC SYSTEMS OF CLASSIFICATION

The first real progress towards a valid system of classification was made by V. V. Dokuchaiev\textsuperscript{2} in 1879. He proposed the following classification:

A. Normal soils (unaffected by other than pedogenic processes).

Class I. Continental humus soils.
(a) Grey northern soils (podsols).
(b) Chernozem (black earth) soils.
(c) Chestnut earths.
(d) Alkaline soils (solonetz).

Class II. Continental swamp soils.

B. Extra-normal soils.

Class III. Denuded or eroded soils.

Class IV. Alluvial and lacustrine soils.

This classification was somewhat modified later. It will be seen that, whilst Dokuchaiev's system marks an advance, in its breadth of application, on purely physical or petrographic systems, it is only devised for the classification of soils of a limited region. No place could be found in such a system for tropical soils or for the soils of Western Europe. The inclusion of extra-normal soils in a system of classification was criticized since they are not actual soils in esse.
F. von Richtofen, in 1886, proposed a system based on the mode of weathering, considered from the physical and topographical standpoint. Against this classification it may be objected that it is simply a classification of regolithic materials.

E. W. Hilgard classified soils broadly into humid and arid, and showed by numerous experimental data the important differences between the groups thus distinguished. This classification, though valuable and valid, did not, however, greatly advance the solution of the general problem.

N. M. Sibirtzev was the first to throw into clear relief the general character of the factors influencing soil formation. According to him, the variety of soils is due to variations in, (a) parent material, (b) organisms participating in the soil forming processes, and (c) physico-geographical conditions, including, above all, climate. The most important factor is the humidity and, in this connexion, Sibirtzev recognized the importance of considering rainfall in relation with temperature in assessing the humidity or aridity of a region.

Three main classes were recognized, namely, A, zonal soils, including the succession from laterite to tundra; B, intra-zonal soils, including salines, peats, and rendzinas; and C, azonal soils, including skeletal and alluvial soils.

The ideas of Sibirtzev, though worked out in one region, Russia, mark a distinct advance towards a world system of classification: yet, with the limited material at his disposal, and with the incomplete knowledge then available as to the constitution of the soil and the nature of the pedogenic processes, it was inevitable that Sibirtzev should lack the perspective possible to later students of the subject.

E. Ramann, in his classification of the soils of Europe, recognized with Hilgard the fundamental distinction of humid and arid soils. His first classification separates the soils in which the weathering is predominantly physical as in arctic and alpine regions from those resulting from chemical weathering. The former class include tundra and
mountain soils. The latter class is sub-divided as follows:—

A. Humid soils.
   (a) Podsolos, sub-divided into geographical groups
distinguished by the degree of podsolization.
   (b) Brown earths.
   (c) Yellow and red earths.

B. Arid soils.
   (a) Chernozems.
   (b) Chestnut earths.
   (c) Grey desert, saline, and alkaline soils.

It is probably mainly to K. D. Glinka7 that we are
indebted for our appreciation of the importance of the profile
in soil studies, since his ideas have exerted the greatest
influence on the present generation of students of the soil.
Glinka's classification is stated in terms of maturity of profile
development and the intensity of leaching by percolating
waters. The following is the classification proposed:—

I. Ektodynamomorphic soils (i.e., soils in which the
external factors of soil formation predominantly
affect the soil character).
   (a) Laterites, red earths, and yellow earths.
   (b) Podsolos, grey forest soils, and degraded
chernozems.
   (c) Chernozems.
   (d) Chestnut earths and their related types.
   (e) Peat soils and mountain soils.
   (f) Saline and alkaline soils.

II. Endodynamomorphic soils (i.e., soils in which the
parent material predominantly affects the soil
character).
   (a) Rendzina, or humus carbonate soils.
   (b) Skeletal soils.

The divisions of the ektodynamomorphic soils are stated
in terms of the character of the leaching which has taken
place (optimal moistening, excessive moistening, etc.).
CLASSIFICATION OF SOILS ACCORDING TO CLIMATE

Since the importance of climate as a factor in soil development has been realized, many attempts have been made to devise schemes for classification of climates, with special reference to their bearing on soil genesis. R. Lang proposes the *rain-factor*, i.e., the mean annual rainfall in millimetres divided by the mean annual temperature in centigrade degrees, as a basis. In its later form, Lang's rain-factor is calculated for the frost-free period of the year. By his recognition that the efficiency of a given annual rainfall as a leaching agency is modified by the temperature, Lang certainly indicates an approximate method for comparing humidities. As limits for the principal soil groups Lang gives the following figures:

<table>
<thead>
<tr>
<th>Soil group</th>
<th>Rain-factor</th>
</tr>
</thead>
<tbody>
<tr>
<td>Peat soils</td>
<td>&gt;160</td>
</tr>
<tr>
<td>Black earths</td>
<td>160-100</td>
</tr>
<tr>
<td>Brown earths</td>
<td>100-60</td>
</tr>
<tr>
<td>Yellow earths, red earths,</td>
<td>60-40</td>
</tr>
<tr>
<td>and laterites</td>
<td></td>
</tr>
<tr>
<td>Saline soils</td>
<td>&lt;40</td>
</tr>
</tbody>
</table>

The inadequacy of this system is evident from the assignment of the black earths to a more humid climate than the brown earths, red earths, and laterites.

A. Meyer has proposed a more accurate measure of humidity in his so-called *N-S quotient*. This is obtained by dividing mean annual rainfall by a coefficient expressing the mean deficit from saturation at the mean annual temperature.

In the following table are given the N-S quotients for the principal soil groups, calculated for the whole year, and for the frost-free period, reduced to the basis of one month.
THE CLASSIFICATION OF SOILS

<table>
<thead>
<tr>
<th>Soil groups</th>
<th>N-S for year</th>
<th>N-S for frost-free period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Desert and desert steppe</td>
<td>0-100</td>
<td>0-5</td>
</tr>
<tr>
<td>Mediterranean</td>
<td>50-200</td>
<td>3-18</td>
</tr>
<tr>
<td>Chestnut earths</td>
<td>100-275</td>
<td>5-10</td>
</tr>
<tr>
<td>Black earths</td>
<td>125-350</td>
<td>8-20</td>
</tr>
<tr>
<td>Brown earths</td>
<td>275-400</td>
<td>18-30</td>
</tr>
<tr>
<td>Atlantic</td>
<td>375-1000</td>
<td>25-80</td>
</tr>
<tr>
<td>Heaths</td>
<td>375-700</td>
<td>25-50</td>
</tr>
<tr>
<td>N. German &amp; Scandinavian</td>
<td>300-1200</td>
<td>20-85</td>
</tr>
<tr>
<td>N. Russian</td>
<td>400-600</td>
<td>20-30</td>
</tr>
<tr>
<td>Alpine</td>
<td>1000-4000</td>
<td>40-350</td>
</tr>
</tbody>
</table>

H. Jenny has shown that the principal soil types in Europe and North America are characterized by definite limits for the N-S quotient.

It is obvious, from an inspection of the figures in the above table, that the principal soil groups are not defined very closely. Indeed, some of the groups distinguished are climatic rather than soil regions. A locality with a reduced N-S factor of 25 might lie in any one of a number of groups. The classification does, however, separate the humid from the arid climates, an end which may be equally well served by Lang's rain-factor.

MODERN GENETIC SYSTEMS

During more recent years, with fuller knowledge of the soils and of the pedogenic processes occurring in different parts of the world, systems have been proposed which aim at the inclusion of all possible types. Thus, D. G. Vilensky distinguishing four broad divisions of soils, based on the dominant factors in their formation. These are (1) thermogenic; (2) phytogenic; (3) hydrogenic; and (4) halogenic.

(1) Soils of the thermogenic divisions are developed in sub-tropical and equatorial regions, in which the dominant factor is the high temperature, which causes (a) rapid chemical decomposition of mineral silicates, and (b) rapid
mineralization of plant residues with production of carbon dioxide. Under these conditions, red and yellow loams and laterites result.

(2) Soils of the phytogenic division are mainly developed in the temperate zones under a wide range of humidity. The dominant factor is the natural vegetation. They are characterized by conditions which favour the accumulation of organic matter in the soil, and involve a less intense weathering of mineral silicates than in the thermogenic divisions. These soils include the chernozem and chestnut-earth groups, the degraded chernozems, and the podsols.

(3) Soils of the hydrogenic division are formed chiefly in cold climates, i.e., in the tundra and the adjacent forest regions. Soil formation proceeds mainly under water-logged conditions with the development of peaty humus. Ferrous compounds such as pyrites, marcasite, and ferrous carbonate, are found in the sub-aqueous horizons. The soils include the tundra soils, the peat podsols, and the meadow soils.

(4) The halogenic division includes soils developed in the presence of sodium salts, and includes the saline, alkaline, and solotli soils.

Intermediate divisions such as thermophytogenic and thermohydrogenic are also distinguished.

D. G. Vilensky, later, developed and re-arranged his system (see p. 445), which throws into clear relief the relative parts played by temperature and humidity in soil formation.

S. S. Neustreuev, in an attempt to classify soils according to the processes by which they have been formed, distinguished hydromorphic and automorphic processes. Hydromorphic processes are those which take place under the influence of ground-water, which directly, or by capillary action, moistens the soil profile. In automorphic processes, ground-water influence is absent.
### The Classification of Soils

#### D. G. Vilensky's System of Classification

<table>
<thead>
<tr>
<th></th>
<th>Arid</th>
<th>Semi-Arid</th>
<th>Medium</th>
<th>Semi-Humid</th>
<th>Humid</th>
</tr>
</thead>
<tbody>
<tr>
<td>Polar</td>
<td>Tundra</td>
<td>Semi-peat soils</td>
<td>Peat and meadow soils</td>
<td>—</td>
<td>Podsolized peat and meadow soils</td>
</tr>
<tr>
<td>Cold</td>
<td>Dry peat soils</td>
<td>—</td>
<td>Black meadow soils</td>
<td>Degraded meadow soils</td>
<td>Podsolized meadow soils</td>
</tr>
<tr>
<td>Temperate</td>
<td>Grey earths</td>
<td>Chestnut earths</td>
<td>Chernozem</td>
<td>Degraded (greyforest) soils</td>
<td>Podsolized soils</td>
</tr>
<tr>
<td>Sub-tropical</td>
<td>—</td>
<td>Yellow soils of dry steppe</td>
<td>Yellow earths</td>
<td>Degraded yellow earths</td>
<td>Podsolized yellow earths</td>
</tr>
<tr>
<td>Tropical</td>
<td>Red soils of semi-desert</td>
<td>Red earths</td>
<td>Laterite</td>
<td>Degraded red earths</td>
<td>Podsolized red earths</td>
</tr>
</tbody>
</table>

The hydromorphous decompositions are divided into two types. In one type, capillary transport is dominant, giving rise to saline soils, and meadow soils with turf or bog iron-ore formation. In the second type, anaerobic conditions dominate profile development and give rise to meadow and peat soils.

The automorphous processes are divided into three types depending on intensity of decomposition of the mineral silicates. In the most intense type, complete decomposition into silicic acid, sesquioxides, and alumino-silicic acids takes place. Differentiation within the profile gives the sub-divisions, laterite, bauxite-laterite, and kaolinite-laterite. A less intense variety of this type is found where intermediate products of decomposition also occur. According to the type of eluviation, red, yellow, and brown earths are produced on the one hand, and podsols and soloti-soils on the other.

In the second type of decomposition, free sesquioxides are not formed and the principal products are acid alumino-silicic acids (clays with high $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio). This type, formed under semi-humid, semi-arid and arid conditions, gives rise to chernozems, chestnut earths, and grey-brown or grey steppe soils.

In the third type of automorphous decomposition, there is only a slight chemical weathering. Products of physical
weathering predominate, whilst the fine material, if present, is mainly of external origin (wind or water transport). The soils produced by this type of decomposition include arctic and alpine soils on the one hand, and desert soils on the other.

A. Stebutt has developed a system of classification also based on the character or tendency of the pedogenic processes. According to this author, the three principal group of processes in soil formations are:—(1) Decomposition, affecting silicate minerals (hydrolysis). (2) Synthesis of new products (formation of the weathering complex). (3) Differentiation into horizons (eluviation and illuviation).

After reviewing these processes in detail, Stebutt distinguishes:—

A. UNDEVELOPED SOIL FORMATION

Class I. Lack of development due to external factors, i.e., low temperature or aridity.

Class II. Lack of development owing to nature of parent material, e.g., quartzite or quartz sand.

Class III. Development potential but unaccomplished, e.g., recent alluvia.

B. DEVELOPED SOIL FORMATION

Class I. Zeolith (Clay) formation.

Sub-class (a) In presence of alkali salts (saline and alkaline soils).

Sub-class (b) After de-alkalination (chernozems, etc.).

Class II. Degradation.

Sub-class (a) Degradation of alkali soils (solot):)

Sub-class (b) Degradation of calcareous soils (degraded chernozems).

Class III. Destruction.

Sub-class (a) Acid: humus destruction.

i. Oxidative conditions.

z. No superficial humus accumulation (brown earths).

β. Formation of peat layer (podsols).

γ. Raw humus accumulations (peat podsols).
ii. Reducing conditions (peats and meadow soils).

Sub-class (b) Destruction by carbonic acid.
   a. Without eluviation (red earths and brown earths).
   b. With eluviation (laterites).

The above scheme gives a comprehensive view of the pedogenic processes and seems capable, by elaboration, of including all possible types and varieties of soil. In view of the importance of the character of the parent material in modifying the trend of the pedogenic processes, systems of classification such as those of Neustreuev and Stebutt mark a decided advance on systems that are stated in terms of climate alone.

The systems hitherto discussed have been based on a consideration of the pedogenic processes. An attempt to classify soils according to the character of the absorbing complex was made by K. K. Gedroiz, who first distinguished, (1) soils saturated with bases and (2) soils unsaturated with bases. The base-saturated soils fall into (a) the chernozem group, predominantly saturated with calcium and magnesium, and (b) the alkaline group in which sodium is dominant, including the saline, alkaline, and soloti soils. The base-unsaturated soils fall into two groups, namely, (a) the podsols and (b) the laterites, including red and yellow earths, varying in the extent to which the weathering complex is decomposed.

Although Gedroiz’s system is valid in itself, it seems scarcely detailed enough in its present form to serve as a world system.

The system of A. A. J. De ’Sigmond, like that of Gedroiz, is essentially a chemical classification. The first division is into I. organic soils; II. chiefly mineral soils, with the co-operation of organic matter; and III. mineral soils. The subsequent divisions are based on the degree of “rawnness” or maturity, the character of the cations associated with the exchange complex, and the degree of leach-
ing. Further subdivisions are possible according to local relief and hydrological conditions.

During the past twenty years, considerable discussion has been directed towards the institution of an agreed world system of soil classification. C. F. Marbut, at the 1st International Congress of Soil Science, held at Washington in 1927, proposed a system of soil classification to include all soils already known or likely to be encountered.

Marbut begins by making a division of soils into pedocals and pedalfers. Pedocals, which may be regarded as synonymous with Hilgard’s arid soils, are characterized by the presence in the soil profile of a zone of calcium carbonate accumulation. Pedalfers, synonymous with humid soils, show no accumulation of calcium carbonate in the soil horizons, but show a differentiation or tendency to differentiation of the clay complex in the different soil horizons, resulting in accumulation of sesquioxides.

In the next stage* of classification, the pedocals are divided into pedocals of the temperate zone and pedocals of the tropical zone, whilst the pedalfers are divided into podsolic soils and lateritic soils. The basis of classification in the second stage is, thus, temperature.

In the third stage of classification, the pedalfers are divided into the following groups: (1) tundra; (2) podsols; (3) brown forest soils; (4) red soils; (5) yellow soils; (6) prairie soils; (7) laterites; and (8) ferruginous laterites. The pedocals are similarly distinguished into, (1) north-temperate pedocals; (2) mid-latitude pedocals; (3) south-temperate pedocals; and (4) tropical pedocals. It will be seen that, in this stage of classification, the distinctions made in the second stage are included.

In the fourth stage of classification, distinctions based on rainfall are used. In the scheme, as outlined by Marbut, this is only applied to the mid-latitude pedocals, giving the succession: chernozems; chestnut earths; brown (desert) earths; and grey (desert) earths. The information at present

*The writer prefers the term stage to category, used by Marbut, and has ventured to introduce this emendation in the present account.
available does not permit a similar differentiation in the case of pedocals of warmer and colder climates. It is considered that this basis of differentiation may be applicable also to the pedalfers.

In the fifth stage of classification, soils are differentiated on the basis of the degree of maturity of the soil profile.

In the sixth stage of classification, distinctions based on the parent material are used, giving the soil series (p. 505), whilst in the seventh stage, the texture is used, giving the type.

**Table XI.V.—Scheme of Soil Classification by C. F. Marbut**

<table>
<thead>
<tr>
<th>Stage 1</th>
<th>Stage 2</th>
<th>Stage 3</th>
<th>Stage 4</th>
<th>Stage 5</th>
<th>Stage 6</th>
<th>Stage 7</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pedocols</td>
<td>Pedocals of Temperate Climates</td>
<td>North-Temperate</td>
<td>Chernozems</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pedocals of Tropical Climates</td>
<td>Mid-Latitude</td>
<td>Chestnut-Earths</td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td></td>
<td>South-Temperate</td>
<td>Brown Soils</td>
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<tr>
<td></td>
<td></td>
<td>Grey Soils</td>
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<td></td>
</tr>
<tr>
<td></td>
<td>Podsolic Soils</td>
<td>Tundras</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Podsols</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Brown Forest Soils</td>
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<td></td>
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<tr>
<td></td>
<td></td>
<td>Prairie Soils</td>
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<tr>
<td></td>
<td></td>
<td>Yellow Earths</td>
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<tr>
<td></td>
<td></td>
<td>Red Earths</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Laterites</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lateritic Soils</td>
<td>Ferruginous Laterites</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Subdivision of categories of Stage 4 according to maturity of profile

Subdivision of categories of Stage 5 according to parent material

Subdivision of categories of Stage 6 according to surface texture
Marbut's classification is shown schematically in Table XLV. It will be seen that spaces are left for the insertion of groups still to be identified and described, so that, with some necessary modifications, the scheme does give a basis for a world-wide system of classification. It will be noticed, however, that no place is found for soils developed under conditions of impeded drainage, such as meadow and vlei soils. Such soils being intra-zonal or azonal, may not be considered as demanding a place in the world-system, on

<table>
<thead>
<tr>
<th>Table XLVI.—American classification of zonal and azonal soils</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soils of the cold zone</td>
</tr>
<tr>
<td>Light coloured soils of arid regions</td>
</tr>
<tr>
<td>Dark coloured soils of arid, sub-humid, and humid grasslands</td>
</tr>
<tr>
<td>Soils of the forest-grassland transition</td>
</tr>
<tr>
<td>Light coloured podsolized soils of the timbered regions</td>
</tr>
<tr>
<td>Lateritic soils of forested warm temperate and tropical regions</td>
</tr>
<tr>
<td>(Halomorphic saline and alkaline) soils of imperfectly drained arid regions and littoral deposits</td>
</tr>
<tr>
<td>Hydromorphic soils of marshes, swamps, seepareas, and flats</td>
</tr>
<tr>
<td>Calomorphic soils</td>
</tr>
<tr>
<td>Fedocals</td>
</tr>
<tr>
<td>Zonal soils</td>
</tr>
<tr>
<td>Pedalfers</td>
</tr>
<tr>
<td>Laterite soils</td>
</tr>
<tr>
<td>1. Tundra soils</td>
</tr>
<tr>
<td>2. Desert soils</td>
</tr>
<tr>
<td>3. Red desert soils</td>
</tr>
<tr>
<td>4. Sierozem</td>
</tr>
<tr>
<td>5. Brown soils</td>
</tr>
<tr>
<td>6. Reddish-brown soils</td>
</tr>
<tr>
<td>7. Chestnut soils</td>
</tr>
<tr>
<td>8. Reddish-chestnut soils</td>
</tr>
<tr>
<td>9. Chernozem soils</td>
</tr>
<tr>
<td>10. Prairie soils</td>
</tr>
<tr>
<td>11. Reddish prairie soils</td>
</tr>
<tr>
<td>12. Degraded chernozem soils</td>
</tr>
<tr>
<td>13. Non-calcic brown soils</td>
</tr>
<tr>
<td>14. Podsol soils</td>
</tr>
<tr>
<td>15. Brown podsolic soils</td>
</tr>
<tr>
<td>16. Grey-brown podsolic soils</td>
</tr>
<tr>
<td>17. Yellow podsolic soils</td>
</tr>
<tr>
<td>18. Red podsolic soils</td>
</tr>
<tr>
<td>19. Yellowish-brown lateritic soils</td>
</tr>
<tr>
<td>20. Reddish-brown lateritic soils</td>
</tr>
<tr>
<td>21. Brown forest soils</td>
</tr>
<tr>
<td>(Braunerde)</td>
</tr>
<tr>
<td>12. Rendzina soils</td>
</tr>
</tbody>
</table>
account of their dependence on special local conditions of topography and drainage. They are, nevertheless, of widespread occurrence and, in the writer’s opinion, should form a major group. Neustreuev, indeed, thus recognizes them in his class of hydromorphic soils.

Marbut’s system has been further developed by the United States Soil Survey workers and the system at present (1948) in use is shown in Table XLVI, which gives the classification of zonal and intrazonal soils. Azonal soils are classified simply with lithosols, alluvial soils, and dry sands.

E. H. Del Villar in his memoir on the soils of Spain and Portugal has proposed a system of classification that is strictly based on the characters of the soil profile and is set out in the form of a “Key,” as follows:

1. Soils developed with free aerobic metabolism
2. Soils developed under conditions of permanent or intermittent waterlogging (hydropedic)
3. Soils with a high proportion of soluble salts.

\[ \text{Saline series.} \]

2

4. Soils with mineral proportions of no soluble salts
5. Soils with a high proportion of the sodium ion in the absorbing complex
6. Soils without this high proportion of the sodium ion
7. Soils with calcium carbonate in the edaphic horizons.
8. Soils without calcium carbonate in the pedological horizons
9. Soils with humus as a fugitive constituent, in which leaching of silicic acid leaves a high proportion of sesquioxides in the profile.
10. Soils with humus accumulation in which leaching leaves a high proportion of silicic acid in the profile
11. Soils with soluble acid humus and marked leaching of sesquioxides from the upper horizons.

\[ \text{Allitic series.} \]

5

12. Soils with mild humus and less intense leaching.
14. Soils with waterlogging, permanent or intermittent, through the presence of ground-water.

\[ \text{Gley series.} \]
This classification can be still further simplified by grouping the eight series into four great cycles:

I. Saline-alkaline cycle.
II. Calcareous cycle (approximately Marbut's pedocals, but with difference of definition).
III. Sesquioxidic cycle: allitic series, siallitic series and acid-humic series (approximately Marbut's pedalfers, but with difference of definition).
IV. Hydromorphic cycle: alluvial and gley series.

Del Villar divides his series in *stages* and phases, the former term denoting a point in the evolution of a series, and the latter term a variant due to causes apart from the characteristic pedogenesis of the profile. Typical stages are *immature, mature,* and *post-mature.* A soil with its natural characters is considered to be in its *primary phase:* the succeeding phases are termed *secondary,* the most important secondary phase being the *agropedic* phase, i.e., the phase resulting from cultivation.

Our knowledge of the soils of the world is probably not yet sufficiently complete for the formulation of a system to include all possible kinds of soil. The following system is put forward tentatively by the writer, as including most of the known soils of the world. It is based on the type of leaching to which the profile has been subjected.

<table>
<thead>
<tr>
<th>Presence of raw humus</th>
<th>Absence of raw humus</th>
</tr>
</thead>
<tbody>
<tr>
<td>Completely leached</td>
<td>Incompletely leached</td>
</tr>
<tr>
<td>(pedocals)</td>
<td>(pedocals)</td>
</tr>
<tr>
<td><strong>Soils with free drainage</strong></td>
<td><strong>Soils with impeded drainage</strong></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Presence of soluble salts</th>
<th>Absence of soluble salts</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Sub-arctic</strong></td>
<td><strong>Temperate</strong></td>
</tr>
<tr>
<td><strong>Sub-tropical and tropical</strong></td>
<td></td>
</tr>
</tbody>
</table>

1. Humus podsols
2. Iron podsols
3. Brown earths
4. Degraded chernozems
5. Prairie soils
6. Yellow podsolic soils
7. Red podsolic soils
8. Tropical red loams
9. Ferrallites
10. Chernozems
11. Chestnut soils
12. Brown desert soils
13. Grey desert soils
14. Tundra
15. Gley soils
16. Gley podsols
17. Peat podsols
18. Peat soils
19. Vlei soils
20. Saline soils
21. Alkaline soils
22. Solotl soils
SIGNIFICANCE OF CLIMATE, GEOLOGY, AND
TOPOGRAPHY

Mature soil profiles are the result of the pedogenic processes acting on different materials, which may be either consolidated rock, or unconsolidated débris, such as alluvium or loess. In arriving at a classification of profiles, it is of importance to obtain a clear view of the relative weight to be assigned to the pedogenic processes on the one hand and to the parent material on the other as factors in soil development.

In the earlier systems of soil classification, the dominant factor was considered to be geology, as affecting the parent material. The studies of the Russian school of pedologists, with their emphasis on climatic factors, have led to greater weight being assigned to the pedogenic processes, which were considered to be capable of obliterating the differences due to parent material. It was found, for example, that chernozem profiles could be developed from parent materials which differed considerably in lithological character. On the other hand, instances were brought forward in which the same parent material gave rise to totally different soils under the action of different pedogenic processes. A classical example is that given by G. Wiegner of two dolerites of approximately the same composition, one of which yielded under English conditions a grey clay, whilst the other, in Southern India, gave rise to a laterite (ferralillite), consisting almost entirely of hydrated sesquioxides.

Since the pedogenic processes are mainly governed by temperature and by the balance of rainfall and evaporation, it is natural that systems of classification should be elaborated that are essentially climatic. Indeed, the criticism might be made that such systems are classifications of climate rather than of soil. Even in classifications, such as that of Marbut, which meet this objection and purport to be based on the actual soil characters, it is impossible to avoid defining the classes in terms of climate.

Now, whilst it is certain that the major differences
between soils are due to the effect of climate operating through the pedogenic processes, the geological factor cannot be eliminated even in the definition of the great soil groups. Many instances may be given in which, under a similar set of pedogenic processes, different parent materials have given rise to soils which must be assigned to different world groups.

B. Polynov\(^2\), in a discussion of the rôle of geology in soil formation and its significance in classification, directs attention to the effect of parent material in modifying the occurrence and distribution of the great soil groups. Quartzose sand enlarges the area of podsol both at the northern and southern limits. Sandy parent materials also increase the desert area. For example, in the Caspian region, whilst loam soils belong to the chestnut earths, sandy soils have a desert character.

In South-Eastern England, typical podsol are developed on the light sands and gravels of the Bagshot Beds. Under the same climate, other parent materials yield brown earths and even soils which may have affinities with terra rossa. It may be said of Britain as a whole that the lightest sands form podsol, whilst parent materials of heavier texture and higher base-status yield brown earths.

In the Transvaal, the weathering of diabase yields red loams which are probably related genetically with the tropical red loams, whilst norite gives rise to black soils, the so-called black turf soils, which resemble, in some respects, vlei soils and must certainly be assigned to a major group distinct from the red loams.

W. H. Bryan and H. J. G. Hines\(^2\) have reported even more striking instances of variety of soil type under the same climate in Queensland. There, within a comparatively small area of approximately uniform climate, soils comparable with the chernozems, podsol, red earths, yellow earths, and laterites of other countries are to be found. This diversity can be satisfactorily explained by considerations of parent material and topographical relief.
The effect of topography was recognized by the late G. Milne, who found in tropical Africa a regular pattern of soils determined by surface relief, with red loams in the higher situations and vlei or mbuga soils in the depressions. He proposed the recognition of a complex of this type under the name of "catena." The catena concept has been further developed by T. M. Bushnell in the light of N. American experience.

Soils derived from calcareous parent materials exhibit certain distinctive features. On the one hand are the red soils of the terra rossa group, whilst on the other hand are the grey or black soils of the rendzina group. It appears that, so long as the calcium carbonate of the parent material can maintain a state of base-saturation, the soil is of a greyish colour with a weathering complex relatively rich in silica and a dark-coloured type of humus. But when desaturation occurs through leaching—a state which may be reached before the complete disappearance of coarsely fragmental limestone—desilicification ensues and reddish or reddish-brown soils are formed, containing free sesquioxides in the weathering complex. This stage is reached most readily in soils derived from hard limestones, which thus tend to be associated with reddish soils, whilst soft limestones generally give rise to greyish or dark-coloured soils of the rendzina group. The ease with which leaching occurs in hard limestone soils, the rapid mineralization of plant residues consequent on their warmth and aeration, all contribute to confer on such soils the characters of soils of warmer climates. Mature leached limestone soils of sub-tropical and even of temperate regions show considerable resemblance to tropical red loams.

It is obvious, therefore, that a classification of climates without reference to geology can give only a partial classification, even of the major soil groups. A map showing the major soil groups must in many cases show different groups occurring under the same climatic conditions.

In the more detailed classification of the soils of a
limited region, geology is frequently the main guide. After all due weight has been given to climatic factors, the most fruitful principle of classification of British soils is the geology, if consideration be given to lithological variations.

Even in a region of uniform climate, geology does not always provide a sufficient basis for differentiation, for important modifications are introduced by topography. Some of these modifications are in a sense geological. The difference between the thin soils of the uplands and the deep colluvial soils of valleys may be expressed in terms of superficial geology. In a detailed soil survey, such distinctions as shallow phase and deep phase are of great importance.

Topography exerts its principal effect on soil development by its bearing on water movements. The same parent material gives rise to markedly different soil profiles according to whether the drainage is free or impeded. With free drainage under humid conditions, the soil is more strongly leached of bases than where the drainage is impeded. This leads to degradation of the clay complex with liberation of silicic acid and sesquioxides. In cool climates, an acid type of organic matter is formed from the residues of vegetation, and a podsol profile results. The brown earths represent an intermediate stage, but seem to be essentially similar to the tropical red loams.

In low-lying situations with a high water-table and impeded drainage, the profile development will depend upon the extent to which the accumulation of acid organic matter occurs. In regions of high rainfall and moderate temperature, a peaty layer is formed and a bleaching of the mineral soil through removal of sesquioxides occurs, resulting in a peat podsol. This process may, indeed, occur in water-logged situations in the tropics.

Where peat is not formed, gley soils occur in cool and temperate climates. These are characterized by grey horizons with rusty mottling in the vicinity of the fluctuating water-table. In similar situations in the tropics, black and
grey soils are found, also with rusty mottling, and even concretionary deposits, in the vicinity of the water-table.

The mineral colour of soils with impeded drainage is generally grey owing to the presence of ferrous compounds consequent on deficient aeration. Further, the low degree of leaching and, in many cases, the relatively high base-status result in the development of a siliceous type of clay complex.

The occurrence of saline and alkaline soils can generally be attributed to the effect of topography operating through drainage conditions.

We have referred repeatedly to climate, geology, and topography as bases of classification. It is necessary to emphasize again that the object to be attained is a classification of soils. The logical order of procedure is first to classify soils and then to attempt correlations with climate, geology and topography. Thus if a study of the soils of a given area of uniform climate shows that they fall into three broad categories, one consisting of soils derived from non-calcareous shales, another of soils from granite, and the third of soils from limestone, it is permissible in describing and defining these categories to allude to the geological parent material, by naming the categories shale, granite, and limestone soils, respectively. Similarly with climate in a region of uniform geology. Climate, geology, and topography, are valuable in arranging the soils of countries, regions, or districts into categories. In an universal system, however, they cannot be so used. The category podsols must cut across geological and even climatic divisions and podsols can be defined only in terms of actual soil profile characters.

The relative weight to be assigned to the factors geology, climate, and topography, will depend on the circumstances of a particular region or country. In Russia, for example, climate has proved the most fruitful principle of classification. In Great Britain, geology gives the clearest clue to soil differences. In every region, however, topo-
Graphy, as affecting the drainage and as modifying the depth of accumulation of the products of weathering, enters as an important factor in classification. The principle to be observed in every case is to classify on the basis of the soil itself, using the profile as the unit of study.

**GENETIC INTER-RELATIONSHIPS AMONG SOILS**

We have already referred (p. 433) to H. Jenny's attempt to express the dependence between soil properties and pedogenic factors by mathematical functions. In a later paper he postulates "five canonical functions of pedology." Denoting the sum of the properties of a soil by $E(s)$,

$$E(s) = f(c, o, r, p, t, \ldots).$$

Where $c =$ climate, $o =$ organisms, including vegetation, $r =$ topographical factors, including hydrology, $p =$ parent material, and $t =$ time. The various sequences are defined as:

- **Climosequences**
- **Biosequences**
- **Toposequences**
- **Lithosequences**
- **Chromosequences**

Expressed in non-mathematical terms this means that an assemblage of soils can be arranged in sequences to show the effect of each individual factor, when all the others are assumed constant. Thus by climosequence he means a range of soils in which the only variable is the climate, and so on. He gives examples from California of a number of sequences, the toposequence being split further into a climosequence (depending on slope) and a hydrosequence, depending on hydrological conditions.

S. A. Wilde (*Forest Soils*, p. 13), developing an idea of Dokuchaiev, expresses the relationship between soil and pedogenic factors as an integral:

$$S = \int (g, e, b) dt$$
where \( g = \text{geological substratum}, \ e = \text{environmental influences}, \ b = \text{biological activity}, \ \text{and} \ t = \text{time}.\)

C. G. Stephens\(^{26}\) has further examined and developed this line of approach to soil relationships. He points out that in nature the variables in Jenny’s equations are not completely independent. Thus—

\[
\begin{align*}
\text{c} & \quad \text{may affect} & \quad \text{o, r, w, p, s, but not t} \\
\text{o} & \quad '' & \quad \text{c, w, p, s, } & \quad '' & \quad \text{t, r} \\
\text{r} & \quad '' & \quad \text{c, o, w, p, s, } & \quad '' & \quad \text{t, } \\
\text{w (hydrology)} & \quad '' & \quad \text{o, p, s, } & \quad '' & \quad \text{t, c, r} \\
\text{p} & \quad '' & \quad \text{o, r, w}, & \quad '' & \quad \text{t, c} \\
\text{t} & \quad '' & \quad \text{c, o, r, w, p, s} & \quad '' & \quad \text{t, c} \\
\text{s} & \quad '' & \quad \text{o, r, w, p} & \quad '' & \quad \text{t, c}
\end{align*}
\]

He gives a complete list of all possible relationships in the form of partial differentials, e.g., \( \frac{\partial s}{\partial c}, \frac{\partial s}{\partial o} \) etc. He considers that the relationship between a soil and its factors is best expressed as

\[
s = f(c, o, r, w, p)\, dt
\]

in which \( t \) is independent and \( c, o, r, w, p \) can have both dependent and independent status.

From a study of published descriptions of certain groups of Australian soils he constructs a number of diagrams showing their genetic relationships and the interaction of the variables.

The relationships between soils of a region are sufficiently involved, and even greater complications and cross relationships may be expected when a larger assemblage of soils is handled. Stephens remarks that if a genetic classification of soils is to be correlated with a morphological classification, the arrangement will need to be on a multidimensional basis. It is therefore hopeless to expect perfect correlation between large scale climatic maps and soil maps, and the same may be said of the other correlations. For the present, soils can be mapped only on their observable morphology and the pedologist must be prepared
to find that the same, or apparently the same, soil profile may be produced by different combinations of pedogenic factors.

PROVISIONAL CHARACTER OF EXISTING SYSTEMS OF CLASSIFICATION

The attempts which have been made to devise a world classification of soils have been encouraged by the impression that identical, or nearly identical, soils can occur in different parts of the earth. Thus, it was generally considered that, in the high plains of the United States, there are soils that can be classed with the chernozem and related groups of Russia. In view of the qualitative character of much of the information relating even to such an important group as the chernozems, it would be safer to regard such correlations as provisional, and to await the accumulation of such a body of quantitative data as shall place their identity or dissimilarity beyond doubt.

This view does not imply that the proposed systems of classification are not worthy of serious consideration. On the contrary, so far as they are based on a considered review of intensely studied soil types, they are of the highest philosophical value. Their provisional character consists in the uncertainty of their application to soils that have not yet been subjected to intensive study, or which have been described only in qualitative and, therefore, subjective terms.

And thus, whilst the various systems proposed may be trustworthy for the soils on which they are based, and suggestive for soils yet to be studied, the immediate task is to enlarge and intensify our information as to the constitution and genesis of the soils of different parts of the world. And if, as is possible though not proven, the soils of the world can be grouped into a finite number of classes, then the modification of existing systems and the synthesis of local classifications will ultimately lead to a world system.
For these reasons, the provisional character of all methods of soil classification hitherto proposed cannot be too strongly emphasized. There are large tracts of the earth's surface where soils have been so imperfectly investigated that no trustworthy conclusions can be drawn as to their relationships. Even for the better known regions, the data are frequently of a qualitative character, and the deductions therefrom may be fraught with serious errors.

Whilst the pedologist should always strive to relate his soils to those already described in other countries or regions, he should not adopt uncritically types recognized under different conditions of topography, geology, and climate. He should, in the first place, distinguish the different profiles in the region for which he is responsible. Each profile should be described with the fullest qualitative and quantitative data, including information as to mechanical composition, structure, base-status, salt content, and the composition of the clay fraction in each distinguishable horizon. In addition, data should also be obtained for the organic matter profile, making use of any methods of fractionation which may become available. With the aid of such data, the profile types should be arranged into a system, which can then be related to systems recognized in other countries.

It is along these lines, in the writer's opinion, that a satisfactory world system of classification will eventually be reached. In some cases it will be at once possible to identify a soil type in one country with a well known soil type already recognized in another country, but the factors of soil formation and the parent materials on which they operate are so diverse that it must be an exceptional case for a soil type isolated, say, in Europe, to be completely identifiable with a type in North America or Australia. Premature identification may be a source of ever-widening error.

There are, therefore, two groups of problems. The one is the philosophical problem of devising a world classification of soils and the other is the practical problem of arranging the soils of a limited area, it may be a parish, a
county, a province, or a country, into a logical system among themselves. We shall discuss this second problem, the problem of the soil surveyor, in Chapter XVIII.

The extension of the broad world classification to include the agriculturally important but philosophically subsidiary distinctions recognized in soil surveys is a task that may well be postponed until more information is available. It may be doubted whether the classification of soils can ever be stated in terms of families, orders, genera, and species, as is possible in the case of animals and plants. Even if it were possible, the distinction between one soil species and another would be far more blurred than in the case of animal and vegetable species.

Finally, one defect of existing schemes may be noticed. In nearly all cases, the ultimate soil-classes are named in terms of colour. Thus, we have red earths, yellow earths, brown earths, and black earths, as well as Russian terms such as podsol, chernozem, and sierozem, which are also colour names. It is obvious that the description "black earth," whether in English or Russian, can apply to soils other than the chernozem \textit{sensu stricto}. Similarly, red earth, merely as a description, can apply to a wide range of genetically distinct soils. If a descriptive nomenclature is to be used, and this seems inevitable in a major classification, it might be better to sacrifice brevity for precision, and to abandon podsol, chernozem, and similar terms as scientific names.

\textbf{BIBLIOGRAPHY}


2 Dokuchaiev, V. V., \textit{Arbeit St. Petersb. Naturforschergesellsch}, 1879, 10; from Glinka, pp. 21-22.


CHAPTER XVII

THE GEOGRAPHY OF SOILS

In the present state of our knowledge, it is, perhaps, hazardous to attempt to give any general view of the soils of the world, for large areas remain pedologically unexplored. Such information as can be given about the soils of these regions is, to a large extent, conjectural, and based on the collateral evidence of climate, topography, and vegetation. Nevertheless, a review of the soils of the world, in the light of such knowledge as we possess, is desirable, and will be attempted, with all reserve, in the present chapter. *

It is only to be expected that fuller information is available in some countries than in others. The lack of accurate data is most marked in tropical regions, where the material hitherto available has consisted mainly of analyses of soil samples made for practical requirements. Only during recent years have descriptions of actual tropical soil profiles appeared in the literature of the subject.

EUROPE.

As may be expected, the information concerning soil is more abundant in Europe than in any other portion of the earth, with the possible exception of the United States; but there are still considerable gaps in our knowledge. A large amount of material is available relating to the soils of Russia, Poland, Scandinavia, Holland, Germany, the former Austro-Hungarian Empire, and Great Britain. Considerably less material exists for judging the soils of Southern Europe.

At the instance of Commission V of the International Society of Soil Science, a soil map of Europe was pre-

*For the sake of convenience in the narrative, the sources are not given in the text, but are set out at the end of the chapter.
pared under the direction of Professor H. Stremme, of Danzig. A provisional map on the scale of 1:10,000,000, with an accompanying memoir, was published in 1927, and a revised map on the scale of 1:2,500,000 in 1935. Making allowance for the somewhat Procrustean system of classification and for the uncertainty as to the position of the soils of certain regions in this system, it is yet possible to make certain generalizations as to the distribution of soils in the different countries.

**Great Britain.** The soils of Great Britain belong mainly to the podsolic group, if in this group we may include the brown earths.* In the provisional soil map of Europe, most of the eastern and southern part of Britain is shown as brown earths, with podsols lying to the north and west. Recent experience has shown the important part played by topography and, above all, by the parent material in determining the character of British soils. The tendency to podsolization becomes more pronounced in moving from south-east to north-west, but quartzose sands in S.E. England carry well developed podsols, whilst, even in the most humid regions of the west, brown earths occur where the parent materials are such as to yield soils of intermediate or heavy texture. Further, the relative immaturity of much of our British soil results in the parent material playing a great part in determining actual soil characters. Age-long cultivation has probably produced considerable modifications, so that soils which were originally podsols or meadow soils have now become brown earths.

Given the importance of geology, there appears to be ample justification for reviewing the soils of Britain in terms of the parent materials which have produced them.

Pre-Cambrian rocks attain their greatest extent in the Scottish Highlands, where they are represented by schists.

*It should be noted that whilst the majority of the well-drained soils of Britain may be described conveniently as brown earths, there is often, particularly in the west, a slight greyness in the surface soil and a development of brownish colour in the subsoil that betokens a certain degree of podsolization. Perhaps the American term "grey-brown podsolic soils" would be more apt.
and sandstones. In the cultivated valleys and foothills, the soils, though generally immature, appear to be usually of a brown earth type where well drained, whilst, in upland areas, podsol are more prevalent. With them are associated gley soils, peat podsol, and peats. The most considerable area of Pre-Cambrian rocks in Southern Britain is in Anglesey, where brown earths are found derived from schists and schistose drifts, with occasional patches of iron podsol in uplands, particularly in association with quartzites.

There is a general similarity between soils derived from igneous and pyroclastic rocks of all periods. Further, with the relative immaturity of British soils, the differences between the weathering of basic and acid rocks are not so marked as in warmer climates. Generally speaking, igneous rocks give rise to brown earths in lowland areas and podsolic soils, sometimes truncated by erosion, in uplands. The soils of the more acid rocks, such as granite and rhyolite, more readily undergo podsolization than those derived from intermediate and basic rocks.

The soils derived from igneous rocks of all periods are generally similar to those derived from crystalline metamorphic rocks. Although, on account of the relative immaturity of British soils, the distinction between the weathering of acid and of basic rocks is not so marked as in warmer climates, differences are observable. Whilst soils derived from acid and intermediate rocks may show varying degrees of podsolization, basic rocks, such as dolerite and norite, yield brown soils of neutral or sub-neutral reaction, showing no evidences of podsolization. We have, thus, from acid and intermediate rocks, a range from brown earths to podsol, the latter predominating in wetter districts and on the more acid rocks such as granite and rhyolite. In upland regions, these podsol may be truncated by erosion. Over basic igneous rocks, we have brown earths only, in varying stages of development.

The sedimentary rocks of Cambrian, Ordovician,
and Silurian age (excepting hard grits, which appear to behave pedogenically as igneous rocks) give rise, in lowland areas, to soils which are generally similar in character and may be classed together as brown earths. But, since they occur mainly in the west under high rainfalls, drainage impedance is frequent and considerable areas of wet soil result. The uplands are occupied by shallow podsols, often truncated by removal of the A horizon so that the surface soil is of a markedly sesquioxidic character. Gley soils, peat podsols, and mountain peats are of frequent occurrence. The soils of the Devonian shales in Devon also may be grouped with those of the older Palæozoic sediments.

The Old Red Sandstone gives rise mainly to brown earths, but some of the quartzose conglomerates carry podsols, as in W. Glamorganshire. The soils of the uplands exhibit various stages of podsolization, modified by erosion. The bright red colours of certain soils of this formation are to be attributed to the effect of the parent material.

The character of the soils of the Carboniferous Limestone is greatly influenced by topography. The calcareous colluvial and glacial drift soils may have affinities with the rendzinas. On the other hand, there are thin leached upland soils which may be assigned to the group of red and brown limestone soils. (See Chapter XV.) In some cases, soil formation is almost entirely absent, and limestone pavements result.

The non-calcareous sedimentary rocks of the Carboniferous system give rise mainly to brown earths, but heavy texture sometimes results in drainage impedance and the development of gley soils.

The Permian formation in England is represented by the Magnesian Limestone, whose associated soils resemble, in many respects, those of the Carboniferous Limestone, and by breccias, marls, and sandstones, whose soils may be grouped with those of the Trias.

The Trias may be considered briefly under the headings, sandstones and marls. The sandstones, which are generally
red in the Bunter, but sometimes brown or grey in the Keuper, give rise to a succession varying from brown earths to podsol. The distinction appears to depend on the actual character of the material, the more quartzose sandstones and the pebbly conglomerates, with their associated drifts, being most apt to become podsolized. Drainage conditions, by their effect on leaching, also play a part. Podosolic characters would, doubtless, be more marked but for the conservative effect of cultivation and manuring on agricultural soils. Gley and peat soils are of frequent occurrence in certain drift areas of Cheshire and Shropshire.

The Triassic (Keuper) marls are mainly reddish clays, sometimes containing calcium carbonate, which, however, is generally leached from the surface horizons. The derived soils are brown earths and gley soils.

Bright red soils occur frequently in association with certain upper horizons of the Carboniferous system, the Permian sandstones, and the Triassic sandstones; but it should be clearly understood that the colour is due to the character of the parent material and is not attributable to contemporary pedogenic processes.

In considering the Jurassic and later systems, with their rapidly alternating succession of sediments, we shall find it convenient to consider them from the point of view of lithology, which, subject to topographical factors, appears to supply an adequate classification.

Soft limestones, such as the Lower Lias and parts of the Chalk, give rise to greyish-brown, greyish, and in some cases, white soils, which may be grouped with the rendzinas. In situations where calcium carbonate is leached away, they may give brown or reddish-brown soils that are essentially brown earths.

Hard limestones, such as the Forest Marble and the Kentish Rag, may give reddish-brown soils akin to the brown earths and, possibly, to terra rossa. Where base-saturation is maintained, the grey or grey-brown rendzina type persists.
Clays, such as the Oxford, Kimmeridge, and Gault clays, generally suffer drainage impedance, and their associated soils may be grouped with gley soils. Horizons of calcium carbonate or gypsum accumulation sometimes occur in the zone of drainage impedance. With free drainage, brown earths are formed.

The sandstones of the Jurassic and later formations are, for the most part, soft or unconsolidated. The derived soils form a range varying from brown earths to podsol according to the lithological character of the parent material. Quartzose sands and gravels, such as the Bagshot Beds, form well-developed podsols with massive humus and iron pans.

The nature of the soils derived from glacial drifts is generally governed by the character of the contributing materials and by the topography and drainage.

Alluvial soils occur in all parts of Britain. They are generally immature, except in the case of old river terraces, where in some cases a certain amount of profile development may have occurred. Dune soils in varying stages of fixation occur in coastal districts.

The most notable development of lowland peat is in the Fen district. Upland and mountain peats occur commonly in elevated areas, particularly in the west.

Much of the soil of upland areas in Britain is very immature and may be described as skeletal. It is probable that erosion has produced considerable modification, so that the shallow profiles encountered are considerably truncated, and represent only the lower horizons of the original profiles existing under the primitive forest cover, which extended up to about 1,500-1,700 feet above the sea level.

*Scandinavia and Finland.* A considerable proportion of Scandinavia is occupied by immature mountain soils with podsolic tendencies. The lowlands of Scandinavia and Finland are occupied by glacial deposits, including boulder clays, lacustrine clays, and fluvioglacial sands and gravels. Podsols, in varying stages of development, form
the typical soil group of these regions and are developed naturally under birch forest, coniferous forest, or in Denmark, heath. Brown earths occur in parts of Southern Scandinavia under deciduous forest. The topography and hydrological conditions associated with a region of glacial drifts result in the widespread occurrence of peats and meadow soils in low-lying situations. In an area of low rainfall in Central Norway, saline soils are encountered.

Iceland. The prevalence of basic parent materials (basalt, liparite, etc.) strongly affects the pedogenic processes. Although as might be expected, Icelandic soils are rich in humus, podsolic soils are not usual. The high base reserves of the parent materials maintain reactions not far below neutrality. From the little that is known of the soils of this island, it would appear that they present a striking instance of the effect of geology in modifying pedogenic processes.

Poland. The soils of Poland show a complicated distribution depending on variations in parent material and in surface relief. Over a large part of the country, chalk occurs covered by loess, which has been partially removed by erosion. Rendzinas occur on predominately calcareous parent materials, chernozems and degraded chernozems on loess, whilst soils varying from brown earths to podsols are developed from loessial debris and from glacial deposits. A considerable area of meadow soils with associated peats occurs in the Priet marsh region of Central Poland.

Switzerland. The soils of Switzerland are dominated by the mountainous character of the country and are mainly skeletal and undeveloped. This applies both to the mountain soils and to the younger alluvia of the valleys. Rendzinas are developed on calcareous rocks in north and central Switzerland, whilst podsols and podsolic soils occur in the southern part of the country, where development is possible.

Germany. Stremme divides the soils of Germany into three zones, namely, the northern coastal zone, the middle loess zone, and the south and south-east mountain zone.
Along the northern coasts, sands and loams are developed from glacial sands and boulder-clays. Forest soil types are also represented. There is a fringe of marshes along the coast and in the west are heath and moorland soils.

In the north-east German plain with a recent glacial landscape, podsolic soils and podsolized forest soils occur principally on the glacial drifts; brown forest soils (brown earths), characterized by a brown B horizon, are also present. The textural types are principally sands and loams, less frequently clays. To the west and south, soils belong principally to the drifts of the older glacializations. The principal soil is the strongly podsolized forest and heath soil, often with ortstein and raw humus. There are also mineral and humous soils of impeded drainage. Among the latter there are numerous areas of high moor peat. Brown podsolized forest soils are not so common. The dominant textural type is sand. It is probable that the soils of this region were originally rich in lime but have now become decalcified. The older drifts are often covered by younger deposits of glacial sands, which are particularly poor as soil formers.

Through Middle Germany there stretches a loess zone with different kinds of soil. In some places chernozem is developed and elsewhere, as in the south-west, degraded chernozems. Where the loess occurs in strong relief, erosion has taken place and undeveloped soils occur. The mountain region in mid-Germany shows a strong influence of topography on soil formation. In the Bunter sandstone region of mid and west Germany undeveloped soils occur on the slopes. Immature podsolic soils occur on acid parent rocks such as granite, sandstone, etc. Even the more basic rocks in this region may give immature podsolic soils. Rendzinas are infrequent because of erosion on the hillsides, whilst in the bottom lands brown forest soils are developed.

In the Upper Rhine Valley are brown forest soils, together with steppe soils. The Upper Rhine plain is mainly sand and loam covered with loess. The celebrated
wine soils of the Rhine Valley are developed from schistose rocks. In the mountains around the Rhine Valley are partly podsollized brown forest soils.

_Czechoslovakia, Austria, and Hungary_. A wide range of soils is shown in these countries. In the mountainous regions, there are immature skeletal soils generally of a podsolic type, but including some rendzinas. Among mature soils of the lowlands the brown earths predominate, but the Central Hungarian plain carries dry steppe soils with associated saline and alkaline (szík) soils. Red soils occur in Moravia and Hungary (nyírok soils) which appear to have affinities with terra rossa but may have resulted from Tertiary weathering.

_Italy and the Adriatic coastal lands_. The soils of northern Italy and the western coastal lands of the Adriatic are mainly brown earths. Alluvial soils occur in the Lombard plain. Elsewhere in Italy, including Sicily, skeletal soils, markedly affected by erosion, prevail. The tendency, however, is towards the development of red soils, which may probably form a transition between the brown earths and the tropical red loams and red earths. Terra rossa and "karst" soils (eroded terra rossa) occur on limestone in the N.E. coastal regions of the Adriatic. In this region salt-bearing soils also occur.

_France_. Podsolic soils are well developed in western, central and south-western France, e.g., in Brittany, the Paris region, and the Vosges, on sands, sandy or gravelly clays, quartz schists, and certain acid igneous rocks. Typical podsols are comparatively rare. They occur under forest (conifers, oak, beech) or heath. Leached soils are widespread. They are often the best agricultural soils. Developed on loess, they form the soil of the great intensively cultivated plateaux of the north-west (Brie, Beauce, Vexin, Santerre). Brown soils are not so common, yet they occupy considerable areas in certain of the regions mentioned (Beauce), also in Alsace, Lorraine, Jura, and in the Midi region. On plateaux, they are generally developed on loess
loams, light or sandy calcareous materials, marls, and sometimes on clays, clay shales, and occasionally on fine sands. They may be very fertile, e.g., the black marl soils of Limagne. The "bocages," grass fields surrounded by woody hedges, so common in the west, correspond with soils of this type developed on rather impermeable parent materials. Elsewhere they are found on slopes. Their agricultural value depends to a large extent on their depth.

On calcareous parent materials, particularly on the sides of valleys, are found rendzinas, generally rather shallow and gravelly. They occupy considerable areas on certain dissected plateaux (Champagne, Poitou, Berry, Bourgogne). They also occur in the warmer regions of the south and south-west.

The red Mediterranean soils, often eroded, occupy parts of the south-east (lower valley of the Rhône).

Recent alluvia are comparatively little developed. The coarser types are poor, but the finer textural types carry rich pasture and may be intensively cultivated. In the coastal regions certain alluvial soils are of high agricultural value, e.g., in Flanders, Brittany, Vendée, and Poitevin.

Skeletal soils and alluvial soils, often of very poor texture, occur in the mountain region of the Alps, Pyrenees and Vosges.

**Holland.** A large proportion of this country is occupied by estuarine alluvium reclaimed from the sea. The soils of the older reclamations may possibly be assigned to the meadow soils and brown earths. Podsol and upland peats occur in northern Holland.

**Spain and Portugal.** In the north and west of the Peninsula the soils are generally of the brown earth or podsol-type. The west central part, for example, much of Algarve (Portugal), Extremadura (Spain), and the foothills of the Sierra Guadarrama is occupied by soils described by Del Villar as "xero-siallitic soils." They are generally developed on acid rock detritus and, although leached, are not markedly podsolized. The natural vegetation was originally
forest, but is now *matorral*, a type of xerophilous scrub. Signs of erosion are everywhere evident. In the eastern part of the country, for example, in Aragon and Murcia, the soils are predominantly pedocalic with local development of saline and alkaline soils and, in the south, of terra rossa. Terra rossa occurs also near Lisbon. The black clays of Andalusia may be rendzinas and are reminiscent of the Houston Clay of Texas. Thin eroded mountain soils cover a large part of the country. Alluvial soils of great fertility occur in the coastal districts of Valencia. Probably no country of western Europe has suffered more than Spain from the deforestation of past centuries. The institution of measures for soil conservation and the right use of the national water resources are essential for the future of Spanish agriculture. Many of the red soils of Spain may be analogous to the red podsolic soils of the S.E. United States.

*Balkan Peninsula*. On account of the strong relief of this region there is generally a marked vertical zonation of soils. Brown earths, red soils resembling the nyirok soils of Hungary, and terra rossa are found in lowland regions, whilst podsolized and skeletal soils, including "karst" soils, occur at higher elevations. Black soils having some resemblance to the black prairie soils of Canada, have been described in the neighbourhood of Sofia. It would appear that considerable modifications have been produced by erosion, particularly in Greece, where the deep stoneless clays of the plains are sharply contrasted with the thin stony soils of the uplands. The tradition of a former cover of soil carrying forest in the uplands of Greece is mentioned in Plato's Critias. Steppe soils occur in northern Greece. The red soils, other than the terra rossas may be analogous to the red podsolic soils of the United States.

*Russia*. Modern conceptions of soil genesis have largely developed from the studies made on the soils of Russia. The dominance of Russian ideas in systematic pedology is the natural outcome of the co-ordinated efforts
of a national school of investigators having as its province the wide range of climatic conditions implied by the vast extent of the Russian Empire.

The succession of soil groups in European Russia is relatively simple. In the extreme north is the tundra region, succeeded to the south by a broad belt of podsols which pass into the degraded chernozems or grey forest soils.

The boundary separating the grey forest soils from the chernozems runs roughly south-west to north-east, passing through the Ukraine. The chernozem zone is succeeded to the south-east by the chestnut-coloured earths, and these, in turn, pass with increasing aridity into the grey and brown semi-desert soils of the Caspian region. Saline and alkaline soils occur as local developments in the Caspian region.

ASIA

India. India may be divided broadly into three main regions, namely, the peninsula in the south, the Indo-Gangetic plain, and the Himalayan Highlands. The most salient feature of the soils of India is the large proportion of alluvium. The greater part of the Indo-Gangetic plain is occupied by the deposits of the Indus, Ganges, Brahmaputra, and their tributaries. Whilst recent alluvium cannot be expected to reflect the operation of the pedogenic processes consequent on the local climate, the older alluvia show varying degrees of maturity. There is thus a marked distinction between the soils developed under the semi-arid climate of the Punjab, and those developed in the more humid climate of Bengal. Soils with carbonate accumulation and soils of the saline and alkaline group occur in the former areas, but are absent from Bengal. A considerable area of desert and semi-desert occurs in Sind and Rajputana.

The southern or peninsular part of India is occupied partly by red soils and lateritic soils and partly by black cotton soils. Soils which may be analogous to the chestnut-coloured earths occur in Hyderabad to the east of the black cotton area.
The black cotton soils, termed regur in Central Provinces and Hyderabad, occupy an area lying to the east of the Western Ghats and including a large portion of these Provinces. The affinities of the black cotton soils with the black earths have been discussed in Chapter XI. Typically they are derived from trap rock—the so-called Deccan trap. The depth varies considerably: it may reach as much as twenty feet in areas of accumulation, but a depth of one to five feet is more usual.

When wet, the black cotton soils form a slimy colloidal mass, but on drying out they readily pulverize, so that the soil is said to "plough itself." The published mechanical analyses of certain typical soils do not, however, show high proportions of clay, and the reason for their highly colloidal character requires investigation.

The development of the zone of carbonate accumulation varies somewhat and may in some cases be almost vestigial. It is possible that the black cotton soils may prove to have affinities with the rendzinas on the one hand and the vlei soils on the other, whilst an analogy with the Transvaal black turf soils may also be suggested. It is significant that in certain areas where black soils occur, the higher ground is occupied by red soils, which may stand in the same relationship to the black soils as the red limestone soils to the black limestone soils in Barbados.

The problem of the mode of origin of these soils and their position in a world classification still remains obscure. They are generally associated with trap as a parent material. Yet trap at higher levels can give rise to lateritic products. Further, black cotton soils can also occur on other parent materials. It seems evident then that they owe their origin to the conjunction of certain types of parent materials with certain topographical and climatic conditions, which have yet to be defined. Affinities with teen-sudas (p. 349) may exist.

Comparatively little information is available as to the red soils of southern India. Developed under a monsoon
climate and mainly derived by primary weathering from crystalline rocks, they are probably similar to the red loams and red earths of East Africa. Laterites are said to occur along the west coast, and in Bihar, Orissa, and Chota Nagpur.

The soils of the Himalayan Highlands show marked vertical zonation and include brown earths, podsolized soils, gley soils, and skeletal alpine soils.

Ceylon. The soils of Ceylon fall into five main groups, namely, (1) Tropical red soils, including lateritic and non-lateritic red loams and red earths, presumably similar to the red soils of southern India; (2) Red soils of the terra rossa type developed on limestone; (3) A group of soils known as "patanas," described as mountain steppe soils. They are sub-divided into dry and wet patanas, the latter occurring at higher elevations (<4,500 ft.). Both are acid in reaction and show a sesquioxidic type of clay. Their position in a world classification seems obscure. (4) Immature soils on Pleistocene and recent deposits; (5) Paddy soils with gley characters, and almost neutral reaction.

Siberia. In Glinka’s "Typen der Bodenbildung" a soil map of the Russian Empire on the scale of 1:20,000,000 is given. The zonation of soil groups in Siberia follows generally the parallels of latitude, and the succession from north to south is from tundra through podsols, black earths, chestnut earths, to grey desert soils. The podsol region extends further south in the mountainous regions adjoining Thibet and China. Most of the area above 60°N latitude is tundra. Degraded chernozems, saline and related soils, and desert soils occur in Turkestan.

Asia Minor and Cyprus. Recent studies have shown the occurrence of terra rossa in Palestine. Comparatively little is known of the soils of Syria and Asiatic Turkey, but they probably include intermediate stages between the brown earths of Europe and the red soils of the tropics. In the drier regions, the black earth to chestnut earth succession probably occurs. Podsols and skeletal soils may be expected in mountainous regions. The soils of Cyprus
include terra rossa, red soils derived from igneous rocks, and steppe soils of pedocalic type.

**China.** Practically all the great groups and sub-groups with the exception of tundra are represented. A large part of the country has been under cultivation for many centuries and the soils are thus considerably modified. Erosion has played a considerable part. China proper may be divided into two main climatic regions, namely, the sub-humid, semi-arid and arid regions of the north and north-west, and the humid regions of central and southern China, and the eastern seaboard. In the former regions, pedocalic soils predominate, ranging from chernozems in varying degrees of maturity to the grey soils of the Gobi desert. Podsoils occur in the north and in the south-west, whilst varieties of brown earths or brown forest soils and degraded chernozems are found in central and N.E. China. Red soils, probably akin to the tropical red loams and red earths, sometimes showing podsolization, are found in the south. There are large stretches of alluvium, which may be subdivided into calcareous, non-calcareous and saline. Patches of saline and alkaline soils are also found. The rice (paddy) soils have been described as ground-water podsoils. Deposit of aeolian material from the loess is an important factor in the maintenance of fertility of cultivated Chinese soils.

**Indo-China.** Tropical red loams and lateritic red earths occur in Indo-China.

**AFRICA**

The soils of Africa lie almost entirely within the tropics and sub-tropics. Analogues with the soils of cool temperate regions are only to be expected in high elevations.

H. L. Shantz and C. F. Marbut, in a study of the vegetation and soils of the continent, have collected a considerable amount of information as to the principal types of soil and their distribution. The map, prepared by Marbut, may serve as a starting point, although it is certain that many
details and, possibly, some of the principal features of it may need modification.

The northern part of Africa is predominantly arid, a large proportion being actually occupied by the Sahara, Libyan, and Egyptian deserts. The soils of the Mediterranean seaboard include pedocals of the chestnut-coloured earth group passing through grey-brown desert soils to desert soils proper. In the wetter regions of the Atlas range, soils of the brown earth type may occur. South of the desert region the humidity increases and there is a transition through chestnut-coloured earths and chernozems to the humid tropical soils of Equatorial Africa. The African pedocals have been little studied except in the Sudan. The soils of the Sudan Gezira, which have been so thoroughly studied by A. F. Joseph and his collaborators at Khartum, appear to be grey-brown soils of the semi-desert, in which the profile has been considerably modified by circulation consequent on the formation of deep fissures during drought.

Pedocalic soils occur also in Kenya, Somaliland, Abyssinia, and Rhodesia. Marbut’s classification of the Transvaal “black turf soils” as chernozems has been controverted by Marchand, and it is possible that equally intensive local study might necessitate further modifications in the provisional grouping suggested.

Saline and alkaline soils occur in association with the alluvial soils of the Nile delta and with the pedocalic soils of the Sudan. Swamp soils occur in many parts along the coast of W. Africa, notably in southern Nigeria.

The more humid parts of the continent are mainly occupied by red soils, which appear to show all gradations between red loams and ferrallitic soils. The latter group appears to be best developed in Sierra Leone, Liberia, and Ivory Coast. Laterites are also widespread.

French North Africa. Here the red Mediterranean soils are the best developed, but are often affected by erosion, which may have been catastrophic in consequence of excessive deforestation.
The lower hills and the plateau of Tell carry sometimes rendzinas, but are mainly occupied by calcareous crusts covered with reddish-chestnut soil. The latter are developed also on old alluvia of the lower plains, e.g., in Atlantic Morocco, the valley of Chelif, and the Plain of Tunis. In these areas, recent alluvia are only slightly developed, but are chemically rich and fertile under irrigation. Occasional patches of saline-alkaline soils may occur.

On afforested hills and mountains, soils vary rapidly with altitude from red Mediterranean soils to podsolic soils. Brown steppe soils and soils with crusts cover the high plateaux. Grey and red semi-desert soils occur in the transition to the Sahara, an immense desert with patches of saline soils, skeletal soils, and other typical desert forms.

Union of South Africa. The great complexity of the parent materials and the resultant configuration of the country caused by mountain barriers, which have a marked influence on the precipitation and other climatic factors, are directly responsible for the great diversity of the irregularly distributed soil types and groups. Topography and resultant climatic features are responsible for humid and sub-humid regions lying adjacent to semi-arid and arid regions, producing corresponding soils. The transitions are generally abrupt, the changes taking place within a distance of a few miles.

Although the soil distribution is irregular, well defined soil groups have, however, been developed. In some instances the parent material is the dominant soil-forming factor that determines soil development and in others again the climate and organic life are the "active" and geology the "passive" factors. Generally, the climatic factor dominates the parent material in the evolution of the soils.

In the arid region of the Union (Karoo), comprising the whole of the Cape Province except for the coastal strip, occur the desert soils, mainly derived from sedimentary rocks. Associated with them are the alkaline or "brak" soils, which have developed under local conditions.
Mechanical disintegration of the parent material, resulting from sudden and extreme temperature changes, predominates over all the other soil-forming factors.

To the north of the desert soils are the Kalahari Sands and Kalahari Sands on limestone. The latter are found on the outskirts of the former in Bechuanaland and South-West Africa.

East of the above-mentioned soil groups are the podsolic soils encircling the Basutoland Mountains and extending eastward to the Indian Ocean. Three stages of development have been observed. The more mature stage has been classified as gley-like podsolic soils.

In southern and central Transvaal and northern Natal have developed the reddish-brown and grey ferruginous lateritic soils. These soils are characterized by either an extremely hard ferruginous hardpan or a thick un cemented concretionary ferruginous layer. They are considered to be "ground-water" laterites.

In the central Transvaal occur the sub-tropical black clay soils, derived from basic igneous rocks—norite and basalt.

On the eastern escarpment of the Drakensberg and Zoutpansberg, running north-south from the Limpopo River to south of Natal, the rainfall and temperatures are very high and here soil groups, classified as laterites, lateritic red, and lateritic yellow earths have developed. These are not ground-water (Buchanan) laterites. To the east of these groups are found the brown forest soils, developed under high temperatures and relatively low rainfall.

The coastal strip of the Cape Province has podsolic and truncated podsolic soils developed from sandstone and shales respectively under winter-rainfall conditions. Skeletal soils are found in Basutoland and other mountainous regions, e.g., Cape Province and Transvaal.

Vlei soils are found in all parts of Africa, south of the Equator, and depend on local conditions of topography. Some examples have already been given in Chapter XIV.
French West Africa. The general distribution of soils is approximately in bands running roughly north and south parallel to the coast. The following is the general succession: sub-desert soils, brown soils, reddish-brown soils under thorn-steppe, leading to tropical ferruginous soils more or less bleached and showing concretions, at first under bush savanna and then under parkland (Senegal, Central and South Sudan, Upper Volta, and southern French Nigeria). Slightly lateritic soils occur under open forest; red soils and very deep grey laterites occur under dense forest.

Lateritic crusts, usually of a fossil character, extend widely between 10° and 13.5° north in the west and between 8° and 13°N. in the east. Generally speaking, they become indurated and develop only after destruction of the forest or bush by fire, followed by erosion of the superficial horizons. Such regions are almost sterile, particularly where the crust forms the surface of the soil.

AMERICA

United States. The great variety of climate, topography, and geology in the United States is reflected in its soils, which include representatives of nearly all the world groups with the exception of tundra and laterite.

The distribution of the main groups is shown in Fig. 19. The most important fact to be noted is the boundary line between the pedocals and the pedalfers (leached soils), which runs approximately north and south through Western Minnesota, Eastern Nebraska, Eastern Kansas, Mid-Oklahoma, and Mid-Texas.

To the immediate east of the great soil boundary from Minnesota to Northern Texas, lie the so-called prairie soils, most typically developed in Missouri. These are considered by the American workers to be a sub-group of the podsols, whilst European workers have considered them to be either meadow soils or degraded chernozems. Interspersed with the prairie soils are areas of meadow soils, as for example, in Iowa.
ZONAL SOIL GROUPS

Fig. 10. Great soil groups of the United States.
Podsols and brown podsolic soils in varying stages of maturity occupy most of the northern portion of the pedalfer region, including Eastern Minnesota, Wisconsin, Michigan, Northern New York, and the New England states.

The Middle-Eastern region is mainly occupied by grey-brown podsolic soils or brown earths, whilst the South-Eastern region from Virginia through the Carolinas, Georgia, Mississippi and Alabama, is mainly characterized by yellow and red podsolic soils with sub-tropical affinities. There are considerable areas of yellow soils in New Jersey and Maryland. These may form a transition between the brown earths and the sub-tropical red soils of the Southern States.

South of the $61^\circ$F isotherm in Alabama, and presumably also in Florida, the weathering is considered by L. D. Bayer and G. D. Scarseth to be of the lateritic type, but it is doubtful if laterite profiles, such as those of the tropics, characterized by superficial incrustations, are present. The relative enrichment of the upper horizons in sesquioxides by removal of silicic acid is, however, well established. It would probably be more correct to describe these soils as ferralitic.

Large stretches of barren sands also occur in the South-East. These, presumably, exhibit podsolization. A considerable area of Florida is occupied by sub-tropical swamp soils. Alluvial soils, sometimes salinized, occur along the coast of the Gulf of Mexico.

West of the great boundary, increasing aridity is reflected in the successive belts of black earths and chestnut earths which pass in the northern region to grey desert soils with associated saline and alkaline soils. The latter groups attain their greatest development in Utah, Nevada, and Southern California. The southern desert soils are of the red and reddish-brown type.

Both in the mountain region and on the Pacific slope, are great areas of immature mountain soils. Brown earths, and, possibly, podsols occur in Oregon and Washington,
Canada. The soils of Eastern Canada adjacent to the United States are mainly podsolic, passing to tundra in more northern latitudes.

The course of the great soil boundary in Canada is somewhat indeterminate, but probably turns westward. Chernozems, prairie soils, and even podsols occur in Manitoba, Saskatchewan, and Alberta. The grey wooded soils of Alberta and Saskatchewan are of particular interest because, although the profile is podsolic in character, they are often of fairly high base-status. Certain soils described as black prairie appear to belong to the meadow soil group. Saline soils occur in Alberta. Great areas of peat occur in the muskeg country of N. Ontario and the northern parts of the prairie provinces.

The soils of British Columbia are largely immature, but probably include brown earths and podsols.

All the groups mentioned pass northwards into the tundra area.

Central America and West Indies. The most detailed study of the soils of these regions is the survey of the soils of Cuba by Bennett and Allison, to which reference has already been made at numerous points in this book. Whilst the general tendency, with the prevalence of humid conditions, is towards the production of a ferrallitic type of soil, all gradations can be observed from the well-known Nipe clay, which consists almost entirely of sesquioxides, to clays such as those of the Truffin and Bernal series in which a siliceous type of weathering complex is present.

Limestones give rise to two groups of soil. On the one hand are the mature leached red soils, exemplified by the Matanzas clay, which are essentially red earths; whilst the unleached limestone soils are represented by the dark-coloured plastic Bayamo clay.

Among the soils of Western Cuba are soils which closely resemble certain series encountered in the South-Eastern United States. They are found in coastal plain areas developed from out-wash material.
There are large areas of poorly drained soils which have affinities with meadow and vlei soils. In certain cases layers of ferruginous concretions or even hardpan (mocarrero) are developed. Soils with layers of carbonate accumulation occur, but it is doubtful if they are to be grouped with the pedocals.

In Cuba, as in other parts of the West Indies, there are considerable areas derived from coral limestone. Some islands, e.g., Barbados, consist almost entirely of coral limestone soils. The contrast between red and black limestone soils has been discussed in Chapter XV.

Many of the soils of the West Indies are very immature, whilst, in other cases, it seems likely that considerable erosion has taken place. There are extensive areas of clay soil in the Naparima district of Trinidad in which it would appear that the cultivated soil is only slightly modified from the parent geological material, which consists of marl or clay. Where maturity can be demonstrated, it is evident that the weathering is of a ferrallitic type. The most characteristic examples are seen in the weathering products of igneous rocks.

The soils of Central America, developed under humid tropical conditions, are generally similar to those of the West Indies, so far as can be judged. Pedocalic soils occur in Mexico, but have been little studied.

South America. Practically all the great soil groups are represented in this sub-continent which includes, tropical, sub-tropical, temperate, and cool-temperate regions. Large areas of alluvial soils occur in the basins of the Amazon, Orinoco, and La Plata. These have not at present been classified, but it is probable that they include considerable areas with developed profiles in addition to the juvenile soils on recent alluvia.

Tropical red soils, including red loams, ferrallites and laterites, occur both north and south of the Amazon basin. Pedocalic soils occur in Uruguay and the northern Argentine. Skeletal soils and podsolic soils occur in the extreme south
of the continent. Skeletal soils occur in the Andes-Cordillera region, with desert and semi-desert soils to the west in northern Chile, and red soils along the north-western seaboard.

*Australia.*—The Australian landscape is largely dominated by soils having recognizable counterparts in other continents. In some of these soils, climate has been the dominant factor in soil formation, in others parent material is co-dominant, whilst in others, the so-called fossil or senescent soils, are seen the results of soil profile development during Tertiary times.

The principal soils of the first or climatic category include: (1) high moor soils of the humid mountain areas in Tasmania, Victoria, and New South Wales; (2) podsols differing from the classic podsols in the minimal development of \( A_{00} \) and \( A_0 \) horizons and maximal development of \( A_1 \) horizons; (3) grey-brown, brown, and red podsolic soils with minimal development of \( A_{00} \) and \( A_0 \) horizons; (4) chernozem-like and tropical black soils; (5) red-brown earths similar to chestnut and reddish-chestnut soils in sub-humid areas; (6) heavy grey and brown pedocalic soils of semi-arid areas; (7) light textured brown and red-brown soils of semi-arid areas with weak pedocalic development; (8) brown and reddish-brown desert soils of the shrub-steppe desert margin, with calcium carbonate and, frequently, gypseous horizons; (9) red sandy and stony desert soils often with "pavement."

The climatic-lithological types include the following: (1) terra rossa on hard limestone; (2) brown solonized soils associated with loess and affected by cyclic salt; (3) red-brown calcareous desert soils on limestone; (4) rendzinas on soft limestone; (5) grey calcareous soils on soft limestone in semi-arid and arid regions; (6) solonchak-solonetz-solotai complexes occurring in southern Australia, where the accessions of cyclic salt are relatively high; (7) red loams of eastern Australia, ranging from Tasmania to tropical Queensland; generally deeply weathered and friable.
The "fossil" soils were developed in Tertiary times on a Pleocene peneplain now elevated to several hundred feet, truncated, and dissected. They are characterized by lateritic horizons, originally developed under the influence of ground-water, and show superficial podsolization. Truncation has generally removed the surface horizons leaving ferruginous crusts or concretionary material at the surface. By dissection the underlying kaolinitic weathering zone may be exposed and a new cycle of pedogenesis inaugurated leading to immature profiles.

Alluvial soils are of limited occurrence, but skeletal soils are common in the mountain regions.

**New Zealand.**—The soils of New Zealand have been divided into the following genetic groups: (1) Recent soils; (2) yellow-grey loams; (3) lowland tussock soils; (4) highland tussock soils; (5) podsols; (6) brown loams; (7) meadow soils; (8) peaty soils; (9) skeletal soils.

Recent soils occupying valley-bottoms are, except where stony or sandy, highly fertile. Yellow-grey loams occur in low rainfall districts in the North Island on parent materials high in lime and, although showing in places a grey A$_2$ horizon, are considered not to be podsols because of their high base content. Lowland tussock soils on the eastern lowlands of the South Island formed under tussock vegetation and low rainfall have a creamy yellow subsoil compacted into a clay-pan at eighteen inches below the surface. The highland tussock soils formed in intermontane basins and plateaux of the South Island under the same vegetation and rainfall as the lowland soils are grey-brown to yellow-brown in the subsoil. The soils of both these groups are little leached and are related to the podsol group. Podsoils next to skeletal soils are the most abundant of North Island soils; in the South Island they occur chiefly on the West Coast. Two kinds are recognized, namely, (a) primary, on acidic volcanic ash deposits; and (b) secondary, on sedimentary beds.

The secondary podsols in the northern part of the North
Island and on the West Coast of the South Island show the deep $A_2$ horizon of the mature stage. In places in North Island the $A_2$ is cemented to a hardpan producing groundwater conditions. Brown loams covering large areas in the North Island are derived from andesite, basalt, and dolerite flows and ash deposits. They are friable soils of excellent physical condition, except in the mature stage of soils formed from basalt rock, which form compact subsoils containing "ironstone" nodules. The brown loams, which are of low silica-sesquioxide ratio are undergoing laterization. The meadow and acid-peaty soils of the tor lowlands, small in extent, have the usual characters of these groups. Skeletal soils on steep slopes—about thirty per cent. of the North Island and more than half of the South Island are shallow and closely related to the parent rock.

EAST INDIES

Our information concerning the soils of the East Indies relates principally to Java and Sumatra; but much of what may be said of the soils of these islands probably holds, mutatis mutandis, for the other islands.

Although Java and Sumatra are entirely within the tropics, generally under a monsoon climate, there is, nevertheless, a marked vertical zonation of soils from the tropical lowland soils to veritable alpine soils. Both sedimentary and igneous rocks occur, and there is, therefore, the possibility of great complexity in the resultant soils.

Most of the soils appear to be relatively juvenile, either through erosion, or else on account of their development from recent volcanic material. The effect of parent material is very marked, but the general tendency is towards the formation of red loams and red earths.

In Middle and Eastern Java, under a relatively dry climate, there are large areas of dark-coloured soils, some derived from igneous rocks, and others from marls and limestones, which appear to have resemblances to the black cotton soils of India.
Alluvial and swamp soils are of common occurrence in the coastal lowlands and may, in some areas, be of a peaty character. Here, as in the uplands, the character of the parent material is frequently of the highest significance in determining the properties of soils, particularly their behaviour under cultivation.

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*The references given here do not include all published works on the distribution of soils, nor, indeed, all the works consulted for the purpose of the above account. A fairly complete list is given, however, in the case of British soils since, in all probability, these will be of principal interest to the majority of readers.

Apart from the specific references given, a considerable body of information relating to soil cartography in many lands will be found in the volume of memoirs on soil cartography collected in 1924 by the late Professor G. Murgoci, of Bucharest, whose generous enthusiasm for the cartography of soils mainly inspired the co-operative efforts of the past decade, and, in particular, the production of the first Soil Map of Europe. Further information will be found in the volume of memoirs on soil nomenclature and classification collected by Professors H. Stremme and B. Aarnio in 1924. Finally, a considerable amount of information relative to the Russian Empire is to be found in the late Professor K. D. Glinka's "Typen der Bodenbildung." For the United States, abundant material may be found in the county soil survey reports published by the United States Soil Survey in cooperation with the different State Experimental Stations.
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CHAPTER XVIII

SOIL SURVEYS

AIMS AND METHODS

WHilst the aims of soil surveys are often mainly of an
utilitarian character, work of this type is of great importance
in elucidating the regional relationships of soils and the
influence of climate and topography on soil genesis. Indeeed, it is only by systematic mapping that the variations in
soil characters within a region can be discovered and related
together into a system, coherent in itself and capable of
being placed in a wider system.

Soil surveys have been in progress for many years in
different countries, but the organization of soil survey work
has been developed more thoroughly than elsewhere in the
United States of America in the form of a national soil sur-
vey under the Federal Government. In other countries,
surveys have been made by workers at provincial institu-
tions, or soil survey work has been included in the activities
of the Geological Survey, as in Germany. In South Africa
and Australia there are soil survey organizations specially
devoted to irrigated and irrigable lands.

It is natural that, in different countries, there should be
considerable variation in the aims and methods of soil sur-
veys. To some extent, the degree of complication en-
countered in the distribution of soils will affect the methods
chosen for their survey. The soil profile, the natural unit of
study, is often so variable in its character, reflecting, as it
does, changes in parent material, climate, and topography,
that the execution of a complete record by mapping all
relevant data may be a very slow procedure. On the other
hand, there are areas even in Britain, where relatively simple
geology and topography under a uniform climate result in extensive tracts of soil which, if not actually constant in character, exhibit variations that can be easily related to topographical conditions.

We have already discussed the broad world classification of soils. In the present state of our knowledge, this classification is still imperfect owing to the insufficiency of existing information as to the soils of large areas of the globe. And, therefore, whilst it is possible, in countries whose soils have been minutely studied, to make direct use of some of the distinctions of a world classification for the purpose of primary study, in other countries the position of their soils in a world system is as yet so uncertain that it is not always possible to assign them to known groups. It is, indeed, certain that fuller knowledge of hitherto unexplored soil regions will add materially to our list of soil groups.

Whilst the survey of the soils of a new country should be carried out with careful attention to relationships with world groups, the first requisite is to obtain such a classification that the soils of the particular country will be thrown into the clearest possible relationship to each other. A valid classification of the soils of a given country is a necessary preliminary to their assignment to positions in a world system.

The scale of actual mapping varies considerably. In intensive surveys of highly cultivated areas, it may be necessary to map on as large a scale as 25 in. to the mile. For ordinary agricultural purposes in this country, a scale of 6 in. to the mile is more convenient. Smaller scales may be more useful for the rapid reconnaissance of large areas. The choice of a scale will, of course, be governed to some extent by the base maps available.

CHARACTER OF DATA COLLECTED

The character of the data collected in a soil survey will vary according to the amount of information already collected by other organizations. In Great Britain, much
information relevant to the soil is already recorded on the Ordnance Survey maps. Fairly detailed information about the geology is generally available, although in many cases the surface geology, which is of principal importance to the surveyor, is rather imperfectly described. In less settled countries, the amount of information available from other sources may be very scanty, and the surveyor is responsible for the collection of all data. In some countries, notably in the United States, aerial photography is used in the preparation of base maps.

From whatever sources they are obtained, the following data appear to be necessary for the characterization of soil conditions.

A. Surface features.
B. Parent material.
C. Soil water conditions.
D. Depth and succession of soil horizons down to parent material with data for each horizon as to:
   a. Texture.
   b. Structure and compaction.
   c. Colour.
E. Nature of cropping or vegetation.

A. Surface Features. Much of the information under this heading is already shown, so far as Great Britain is concerned, on the Ordnance maps used as base sheets. Where, however, contour lines are at wide intervals in gently undulating country, important surface features may be unrecorded on the base maps and should be shown on the soil map. It is convenient to distinguish a small number of categories for record. In Great Britain, for example, the distinctions, flat, rolling, steep, very steep, and broken or irregular are noted by appropriate symbols. Arrows are inscribed on the maps to show directions of slope. Special
symbols may be used to denote broken country, knolls, and depressions. The character of the micro-relief is often of great importance.

B. Parent Material. Here it is necessary to insist again on the distinction between parent material and parent rock. In some cases, as in the case of the deeply weathered layers often found in the tropics, the parent rock may be many feet below the surface and below the range of operation of pedogenic factors. In the case of glacial drifts and alluvia the parent rock may be remote and the period of weathering equally remote in time. In other cases the parent rock may itself be within the range of the pedogenic processes and may have an effect on the moisture régime and upon the growth of crops. This is the case with many shallow upland soils. In immature soils, the soil profile may consist entirely of slightly modified parent material without true soil horizons. This is the case in recent alluvial soils such as those of recent reclamation in Holland. There are also other instances in which the actual cultivated soil would appear to be almost unmodified parent material. These may be encountered in regions with rolling topography where erosion has removed the original soil horizons from unconsolidated parent materials. The agricultural soils in such cases are, pedologically speaking, C horizon material. The writer would assign certain sugar cane soils of the Naparima district of Trinidad and many of the Old Red Sandstone Marl soils of S. Wales and W. England to this class.

The soil profile is developed from more or less weathered rock and it is this weathered and, in some cases, transported material that forms the parent material of the soil. In some cases, as when a shallow soil is developed in decomposing rock, the rock may be thought to form the parent material. It is better, however, to bear the distinction in mind and regard the soil as being formed not directly from the original rock but from its weathered products.
It is often a matter of difficulty to decide the depth to which the soil horizons extend and the division must in some cases be arbitrary. Even in soils derived from material weathered in situ, there may be a thickness of weathered material overlying the parent rock which does not belong strictly to the soil profile, but which should be considered as parent material from which the soil profile has been differentiated. This is particularly the case in many tropical soils.

Although the basis of soil classification is the actual profile, it is convenient to be able to associate a particular soil with a given parent material and even with a given parent rock where the other pedogenic factors remain constant and the only variable is the parent material or the parent rock. The parent rock is not so much a definition of the soil as an indication where that soil may be expected to occur.

In classifying parent materials the lithology of the parent rock is of greater importance than stratigraphy. This means that it is necessary to take account of lithological variations in rocks of the same age. It means also that rocks of different age can be grouped together if they are lithologically similar. A classification of parent rocks is essentially lithological, but must also take into account conditions of weathering and transport.

C. Soil Water Conditions. Both from the standpoint of soil classification and of practical utility, distinctions under this head are of the greatest importance. It is convenient in the first place to distinguish soils of free drainage from soils with impeded drainage. The former may be subdivided into soils that are satisfactory and soils liable to drought. Soils with impeded drainage may be divided into soils with positional water-tables as in the valley bottoms or coastal regions, and soils of elevated regions having impervious strata. Varying degrees of wetness may be distinguished and, in addition, it should be noted whether wetness is continuous or intermittent.
It sometimes happens that excessive wetness may alternate with excessive dryness in the same soil. Changes in level of the water-table in a coarse sand may lead to such alternations. Further, in heavy clay soils, winter wetness may have the effect of restricting root development to the surface horizon. When drought occurs, the moisture within root range is quickly exhausted and capillary rise is too slow to repair the losses by evaporation and transpiration. These circumstances are generally reflected in the character of the natural vegetation or, in cultivated lands, the grass herbage. Indeed, natural vegetation or grass herbage are often the surest clues to the nature of the soil.

D. Depth and succession of soil horizons. The unit of soil study is the complete succession of horizons down to the parent material from which they have been differentiated. The soil survey should, therefore, include descriptions of all types of soil profile likely to be encountered. The data are conveniently recorded in notebooks with corresponding references to the field map. The actual description should be made from the examination of trenches or pits and uniformity of profile characters checked from point to point by auger borings. The actual depth of exploration will vary somewhat, but should reach to the full extent of the soil horizons. It is not generally necessary in Britain to go below three or four feet, but it is desirable to carry the examination to greater depths at intervals for confirmation. Profile exposures at quarry faces or gravel pits should be used with caution as they generally represent abnormal conditions.

The first step is to recognize the successive soil horizons, The approximate depth of each is recorded, together with data for texture, structure, and colour.

a. Texture. For some purposes, distinctions in texture are of great importance. For ordinary soil surveys, however, it is not practicable to have more textural grades than can be distinguished in the field by personal judg-
ment. In Great Britain, the following texture grades are recognized:

<table>
<thead>
<tr>
<th>Texture Grade</th>
<th>Code</th>
</tr>
</thead>
<tbody>
<tr>
<td>Light Sand</td>
<td>Sa</td>
</tr>
<tr>
<td>Heavier Sand (or Loamy Sand)</td>
<td>Sb</td>
</tr>
<tr>
<td>Light Loam</td>
<td>La</td>
</tr>
<tr>
<td>Medium Loam</td>
<td>Lm</td>
</tr>
<tr>
<td>Heavy Loam</td>
<td>Lb</td>
</tr>
<tr>
<td>Clay</td>
<td>C</td>
</tr>
<tr>
<td>Coarse Silt</td>
<td>Za</td>
</tr>
<tr>
<td>Fine Silt</td>
<td>Zb</td>
</tr>
</tbody>
</table>

Intermediate grades may also be distinguished, such as sandy clay (Cs) and clayey sand (Sc).

In the United States, a much larger number of textural grades is recognized. The judgment as to texture obtained in the field is checked from time to time by actual mechanical analyses in the laboratory. With experienced surveyors, the correspondence is fairly close.

Special attention should be given to the presence of a texture profile.

b. Colour. Colour is perhaps the most difficult of all soil properties to describe accurately. Up to the present, no generally accepted scheme of colour description has been agreed upon. Such a scheme would be of considerable value, for there is, perhaps, no other soil property which is of such value for diagnostic purposes in the field. At present, the recognition and recording of colour distinctions is generally subjective and personal. It would be helpful to field workers if a select colour atlas were generally available for use in the field. The colours might be chosen to cover the range likely to occur in the field. Such a colour atlas is now actually in use in the United States Soil Survey.

The most important distinctions to be noted in soil surveying are between grey colours, on the one hand, and red and brown colours, on the other. Grey colours are found in arid soils, in the bleached layers of podsol soils, in soils with impeded drainage, and, in Britain and Northern France, in immature or eroded chalk soils. Brown and red colours are found in tropical soils and also in soils of temperate climates.
formed with complete leaching in the absence of acid humus. Red colours may be conferred by the parent rock as in the soils of the Trias and Old Red Sandstone in England. Experienced surveyors can frequently detect the presence of admixtures of drift from the colours of soils. The presence of rusty-brown, orange, or yellow streaks or mottlings should be particularly noticed, since they are valuable evidence for the occurrence of waterlogging or drainage impedance.

c. Structure. This soil character has been given particular attention by Russian and American soil workers. In Britain, it has probably not been given sufficient importance. This may be due to the fact that British soils are generally moist, in which state structural characters are not so readily observed. Some of the typical structures recognized by workers in the United States are shown in Plate III. In addition to the macro-structure implied by the terms, granular, nut, platy, etc., it is desirable to consider also the micro-structure, noting the character of the small crumbs or fragments and such appearances as the presence of pores.

In connexion with structure, note should be made of changes in the degree of induration or compaction in the different horizons.

E. Natural Vegetation and Cropping. In a new country, a record of the natural vegetation is of the utmost importance, and may provide important clues to the distribution of soils. In difficult country, aerial photography has proved to be a convenient method of mapping vegetative cover if the aerial maps are compared with accurate ground surveys of selected areas.

In settled countries, the record of cropping at the time of survey is only of fugitive significance, but may be important in connexion with advisory work. In Great Britain, data on grass herbage are of considerable value both from the practical standpoint and also as an aid to the recognition of certain soil characters, more particularly the base-status and the moisture relationships.
Where possible, notes on weed and hedgerow flora should be made. Tree growth, also, can often be correlated with soil characters.

In the uncultivated lands of Wales, an exceedingly close correspondence is found between soil characters and vegetation type. By a view of the vegetation alone, it is frequently possible to obtain a fairly correct idea of the distribution of the different classes of soil.

G. R. Clarke has summarized with explanatory notes the actual procedure followed in the field study of soils. The same author has also prepared a Field Handbook and a form for the recording of profile data.

**EFFECT OF SUPERFICIAL GEOLOGY AND TOPOGRAPHY**

The problems confronting the soil surveyor will vary with the region to be surveyed. Whilst, in some areas, soil differences are easily seen, and the different classes recognized can be readily correlated with geology and topography, in other areas, the changes in soil character are so frequent that a trustworthy map can be produced only by the most minute examination. Complications, in such cases, generally reflect the intricacies of superficial geology. This obtains particularly over large areas of England and Wales, where the main task confronting the surveyor is the perception and delineation of glacial, colluvial, lacustrine, and alluvial deposits. Considerable help can often be obtained by an intelligent and imaginative interpretation of topography; yet there are also areas, such as certain alluvial and lacustrine flats, where the facts can be ascertained only by closely ranged auger examinations. In such cases, it may be impracticable to insert all the existing detail, but a close survey of specimen areas may serve to indicate the kind of variation to be expected over the area as a whole.

**CLASSIFICATION INTO SERIES**

A principal object in soil surveys is the production of soil maps in which the information gathered in the survey is shown pictorially. Unless the surveyor is to use the same
almost infinite range of colours used by an artist in painting a picture, he must have a limited number of categories for representation on his map. It may be, of course, that in thus limiting himself he is oversimplifying nature, but the error may not be serious if the limitations of soil maps are recognized and appreciated. For example, the clear, firm boundary lines of soil maps rarely occur in nature. At an early stage in the work, the soil surveyor must decide on his legend, i.e., the categories to be recognized and the colour or pattern by which they are to be represented on the map.

The arrangement into categories is based on the profile characteristics. Soils with similar profiles derived from similar material under similar conditions of development are conveniently grouped together as a series. A series is further sub-divided according to textural variations. In the United States of America, where this system is used in the Soil Survey, series are named after the localities where they are first studied or where they attain considerable development. The textural distinctions are shown in the types. Thus, the Leonardtown series include such types as the Leonardtown medium loam, the Leonardtown fine sandy loam, and so on. Finally, types may be subdivided into phases, depending on depth, slope, and degree of stoniness.

The series system with some adjustments has been adopted in Great Britain. In the early stages of the soil survey of Wales, it was found convenient to make a preliminary classification into suites, i.e., soils derived from the same or similar parent material. The qualification "or similar" was introduced because it was found that parent materials of different geological age, but of generally similar lithological character, give rise to the same type of soil profile under similar conditions of formation. The soils of a suite are divided into series according to the mode of development, as reflected in the profile. For example, the Powys suite comprises soils derived from weathered Cambrian, Ordovician, and Silurian non-calcareous hard sedimentary rocks,
excluding crystalline grits. The following series have, up to the present, been recognized:—

(1) Powys Series. Well drained soils, developed from sedentary parent material on the parent rock.
(2) Penrhyn Series. Well drained soils derived from glacial drift or colluvial material.
(3) Cegin Series. Soils with impeded drainage derived from glacial drift or colluvial material.
(4) Conway Series. Alluvial soils.
(5) Hiraethog Series. Podsol:s developed in sedentary parent material or drift.
(6) Cymmer Series. Eroded soils resulting from the truncation of podsolic profiles.

Each series gives types according to texture. Whilst, theoretically, a series might include a complete textural range, the variation is comparatively small in practice.

In the early stages of a survey, it is generally desirable to avoid too much restriction of the number of series. It is safest to assume, in the first instance, that distinct parent materials, even though lithologically similar, give rise to distinct soils. When, however, further study shows that the soils are indistinguishable, the series may be combined.

For example, in the author's experience, there is no essential difference between the soils from similar Cambrian, Ordovician, and Silurian sediments, and they are therefore grouped together. If it can be demonstrated that soils derived from Devonian or Carboniferous shales are essentially similar to these, then they also can be included in the larger group. The criterion is always the soil profile.

The suite as a higher category than the series was proposed by the present writer. It was, perhaps, not entirely satisfactory because, at the time of its introduction, the distinction between parent material and parent rock was not sufficiently appreciated. More recently, other workers have proposed categories which, whilst yielding groups essen-
tially similar to suites, are more in accord with modern pedological doctrine. The most widely accepted grouping of soils for mapping purposes is the catena first proposed by G. Milne† and further amplified by T. Bushnell§.

Milne's catena is essentially a group of soils derived from similar parent materials occurring in a complex in which the individual members are differentiated by topographical and hydrological conditions. For example, in E. Africa, there is often an alternation of red loams or red earths in elevated well drained sites with vlei or mbuga soils in depressed sites with seasonal drainage impedance.

The catena concept, apart from its value in directing attention to the relationships of different soils derived from similar parent material, is a convenient mapping unit in areas of great complexity so intricately arranged that detailed mapping would demand too great an expenditure of time and effort. Where catenae are mapped as units, it is necessary to work out the detailed distribution in a typical succession in order that the significance of each catena may be appreciated.

A similar concept to that of the catena is the association, introduced in Canada by J. H. Ellis and W. H. Shafer&, and used by R. Glentworth in the Soil Survey of Scotland.

In planning a soil survey, it is always a matter of difficulty to decide what degree of detail should be attempted. The standard scale in Great Britain is that of 6 inches to the mile, but for some purposes a scale of 25 inches to the mile might be desirable. Indeed there is scarcely any limit to the amount of detail that can be inserted, given sufficient time. In some areas of great complexity, a ten-acre field might occupy an experienced surveyor for many days if each small variation in depth, texture, slope, and moisture conditions were shown. At the other extreme is the temptation to produce soil maps quickly, partly to satisfy the demands of practical men and partly to impress those who distribute grants in aid.
In a country like Great Britain, a reasonable aim would be the production of maps covering the whole of the country on the scale of 1 inch to the mile without undue delay. In order to produce such maps it would be necessary to map sample areas in detail and then proceed with reconnaissance mapping on the scale of 2½ inches or 1 inch to the mile over larger areas. In reconnaissance mapping, the unit might be the catena or association, but no area should be mapped as covered by a given catena unless the degree of complexity has been discovered and exemplified by the mapping of sample areas. The first soil map of the country would thus be a semi-reconnaissance map reinforced by well-distributed areas of detailed mapping.

Areas of mountain and waste present special difficulties. The base maps are generally lacking in detail, yet the actual soil pattern can often be exceedingly complicated. Here it might be sufficient to recognize a limited number of soil complexes and to illustrate them by detailed maps of sample areas.

An important part of a national soil survey is the correlation of the work of different surveyors. It is undesirable that the same soil series should be mapped under different names in different parts of the country; yet it is equally important that two series should not be assumed identical without careful comparison of the field and laboratory data. In the early stages of a survey, it is probably better to err by having too many than by having too few series. Frequent visits by surveyors to each other's areas are helpful.

THE USE OF LABORATORY DATA IN SURVEY WORK

The first appraisal of the soils of an area is, of course, made in the field and, in the opinion of some surveyors, soils should be mapped on field characteristics. Yet the identity of a given soil profile must rest ultimately on some securer basis than qualitative and personal judgement. It is therefore desirable that representative profiles should be submitted to careful laboratory examination and that the
series and type characters should be defined in a quantitative manner.

A complete definition of a series should give data on the following points:—

1) *Mechanical composition of the principal types.* Whilst a certain textural range within the series is admissible, it is often the case that the mechanical composition curves show a general similarity that suggests their relationship. Mechanical analysis should always be used as a check on field estimates of texture.

2) *Organic carbon and nitrogen content.* These figures will show some variation, yet the vertical variation of the organic carbon content within the profile may prove to be characteristic of a series. For example, soils with impeded drainage frequently show a sharp decrease in organic carbon content at the horizon of impedance.

3) *The composition of the clay fraction.* Ideally this should be given by a determination of the relative proportions of the different clay minerals present, together with data, such as those given by the Tamm method for the more mobile constituents, in particular the iron oxides. The gross chemical composition of the clay fraction as expressed by the $\text{SiO}_2/\text{R}_2\text{O}_3$ and $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio, although less satisfactory, can often give useful indications and, in the absence of more significant data, should always be determined.

4) *The base-status.* The variations in the base-status, as shown by the content of exchangeable bases or by the pH, are of great diagnostic value. In cultivated soils, however, liming may produce modifications, of which the surveyor should be aware.

5) *The nature and distribution of soluble salts.* Whilst these do not occur to any marked extent in British soils, there are countries such as the Sudan, S. Africa, and Australia, where information on this point is of the greatest value in characterizing and classifying soils. Determination of exchangeable sodium may give valuable information.
(6) Mineralogical composition. A knowledge of the unweathered minerals present in a soil is of great diagnostic value, provided the mineralogical composition of the parent material is also known. R. Hart, in Scotland, has found the minerals of great assistance in identifying glacial drifts. Further, under tropical conditions, a knowledge of mineralogical composition may be useful in estimating the extent to which the soil has arrived at the end-point of its weathering. Unfortunately, available methods of analysis, in addition to being very laborious, do not succeed in giving a complete picture of the mineralogical composition of the soil, for they are not applicable to particles of finer grade than fine sand.

(7) Physical properties. It would be desirable to be able to express quantitatively those physical properties of a soil which affect its water-air relationships. Data showing the variation in pore-space throughout a profile would be exceedingly valuable as a measure of the degree of compaction. Unfortunately, the physical analysis of soils is at present in a less advanced stage of development than the mechanical and chemical analysis.

We have already alluded to the importance of obtaining a satisfactory scale for the expression of soil colours. When such a scale is available it would be desirable that colours should be recorded both for wet and dry soil, and after removal of humus by peroxide treatment.

SOIL MONOLITHS

In connexion with soil surveys in Russia and the United States, it has been found helpful to preserve records of the principal kinds of soil profile in the form of soil monoliths. These are actual columns of natural soil to the full depth of the profile. The taking of soil monoliths is an operation demanding considerable ingenuity. Essentially, the procedure consists in excavating a pit to the full depth of the soil profile and, after cutting away three sides of the proposed monolith, fitting a stout box. When the monolith
Fig. 20. - Portion of six-inch soil map of Anglesey, XXI, N.E. (From "Journal of the Ministry of Agriculture." By permission.)

Surveyed by D. O. Hughes and J. C. Jones.

The arrows point down slopes.

fits snugly in the box, the fourth side is then cut away. The cut face is then chipped to obliterate spade marks and to throw up as far as possible the natural structure. The greatest difficulty is experienced in stony soils.

As a substitute for monoliths it has been proposed to apply a flexible material covered with an adhesive substance, such as glue, to a profile face. When adhesion is attained, the material is pulled away with the adhering soil.

PREPARATION OF MAPS

In soil survey work in Great Britain, the Ordnance Survey map on the scale of six inches to the mile proves the most convenient base map for use in the field. On it should be recorded all data which can be conveniently inserted without undue crowding. In complicated areas, however, it may be desirable to insert key-letters or numbers referring to descriptions made in a field note book. Boundaries between soil series should be drawn in the field.

It is, perhaps, a matter for debate whether type boundaries should always be inserted. Generally speaking, the textural range within a soil series is not wide, and the delineation of fine textural distinctions may involve an amount of labour incommensurate with the value of the information given. In such cases it should be sufficient to insert textural data (La, Lb, etc.) at frequent intervals over the extent of each series on the map. In intensively cultivated areas, textural details may be of sufficient importance to warrant their close study and accurate delineation.

An example of a completed soil map of a part of Anglesey on the scale of six inches to a mile is shown in Fig. 20. The reproduction, for publication, of six-inch soil maps would generally be rather too expensive and they should be regarded mainly as the repositories of the information gathered in the field work.

For ordinary publication it is desirable to make a generalized map on the scale of one mile to the inch. Such a map would simply show the soil series and would not ordinarily carry detailed information as to textural varia-
Fig. 21.—Soil series map of portions of Anglesey and Caernarfon.
tions within the series; nor would it be practicable to show on such maps the minor topographical details shown on the six-inch maps. An example of such a generalized map is shown in Fig. 21, in which the area shown in Fig. 20 is indicated by the small rectangle. It should be added that both maps have been further reduced for reproduction; but the difference in the amount of detail shown will be evident.*

LAND CLASSIFICATION

By this term is understood any type of classification that distinguishes land according to its use or potentialities. Whilst in a soil survey or a vegetation survey, the objective is simply the recording of physical facts, in a land classification an attempt is made to go a stage further and to draw certain conclusions from the physical facts taken in relationship to any other relevant non-physical factors such as economic conditions, sociology, and the like.

A very common type of land classification is according to “use-capabilities,” and C. E. Kellogg* defines this as “one in which bodies of land are classified (on the basis of physical, or upon the basis of both physical and economic considerations) according to their capabilities for man’s use, with sufficient detail of categorical definition and cartographic expression to indicate those differences significant to man.”

Now whilst physical facts relating to soil, topography, and climate are objective and ascertainable even although there may be different methods of arranging the observed data, when land use and potentialities have to be considered, the resulting classification contains a large subjective element and is not purely scientific. For example, all experienced pedologists might agree that a certain soil is a grey-brown podsolic light loam and might agree as to its depth, slope, and moisture conditions. It would not be so

*Note.—Since the execution of these maps the Anglesey suite has been merged in the Bangor suite. The former Gaerwen and Ebenezzer series become the Arvon series, the former Gesail and Sion series become the Eivion series, and the former Braint and Glanadda series become the Ynys series.
equally evident that this soil is suited to a particular type of agricultural use. Even in the apparently simple task of classifying soils according to their productivity there is ample latitude for difference of opinion, according to the meaning to the assigned to the term productivity. And therefore, although for some purposes it may be permissible to prepare maps showing land set out in different categories of land use or potentiality, it should be realized that such maps have much less objective value than well executed geological or soil maps.

Land classification maps should be designed to answer specific questions. The more general the question, the less accurate is the map likely to be. For example, for some purposes there may be a demand for maps showing the different agricultural values of land. Such a map in its simplest form would show, perhaps, all the area divided up into good, medium, and poor land. The object of such a map might be to serve as a guide to planners who wish to avoid the diversion of the best agricultural land to urban or industrial use. Unfortunately, there is no clear and unambiguous definition of agricultural value or even of soil fertility, and the map produced must depend on a personal interpretation of the effect of the known physical factors.

With more clearly defined objectives greater significance is possible. For example, it might be possible to attain a fair degree of significance in maps showing liability to erosion or suitability for the growth of particular crops such as potatoes or market garden crops.

In spite of the difficulties of integrating all the site and profile characters of a soil to give a single index of productivity or fertility many schemes have been proposed for solving this problem. An early example of this is R. E. Storie's index, developed in California. Storie uses three groups of factors, namely, (1) general profile characters, (2) texture, and (3) conditions such as drainage, acidity, salinity, etc., modifying the productivity of the soil. A mark is assigned to each of the three groups of factors expressing
a percentage of optimum. The three are then multiplied together and expressed as a percentage of the maximum. This is claimed to give a fairer appraisal than addition, but it should be noted that the multiplication together of three moderate percentages, e.g., 70%, will give the very low final figure of 34%.

More usually, land is arranged into groups according to increasing or decreasing agricultural value. An example of this is the classification devised by L. D. Stamp\textsuperscript{11} for the Land Utilization Survey of England and Wales. In this system, land is divided into 10 major categories, which are grouped into Good (4 categories), Medium (2 categories), and Poor (4 categories). The categories within the groups are based on qualitative distinctions such as suitability for grass or arable. A Committee of the West Midland Group for Post-War Reconstruction and Planning\textsuperscript{12} devised a modification of this system, whereby the three groups were defined in terms of soil and site characters. The major categories are thus defined:

**Good quality land.**—Highly productive under good management. Land in this category has the following characteristics:

\[
\text{Site \{ Not too elevated, level, gently sloping, or undulating. Favourable aspect. }
\]

\[
\text{Soil \{ Deep, with favourable water conditions (actual or potential). Texture—mostly loams, but including some peats, sands, silts, and clays. }
\]

**Medium quality land.**—Land of only medium productivity even under good management. Productivity limited by reason of unfavourable operation of one or more of the factors of site or soil character, e.g., by reason of—

\[
\text{Site \{ High elevation. Steepness. Unfavourable aspect. }
\]

\[
\text{Soil \{ Shallowness. Defective water conditions. }
\]
Poor quality land.—Land of low productivity by the extreme operation of one or more factors of site and soil.

Another approach to the problem of land classification has been through "productivity ratings" based on observed performance of crops and grassland. The method is based on the work of J. K. Ableiter. In the county reports of the United States Soil Survey, figures are given for each type showing the percentage of the "standard" yield obtained for each crop under (A) average farming and (B) the best type of husbandry. The productivity grade for a given soil is obtained by weighing the percentage yields according to the regional importance of the crops. Soils with a weighted percentage of over 90 are rated as Class I., those between 80 and 90 as Class II., and so on. The difference between the A and B ratings gives a measure of the response to good husbandry.

The problem of land classification transgresses the boundaries of soil science and enters the domains of economics and sociology, where facts are ascertainable with much less precision than in the natural sciences. Probably the best service that the soil surveyor can render to the would-be planners is to provide accurate soil maps and clear explanations of their significance.

G. V. Jacks has recently summarized the present state of the problem of land classification. The reader may be further referred to the Proceedings of the 1st National Conference on Land Classification (1940), published as Bulletin 421 of the University of Missouri under the title "The Classification of Land."

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CHAPTER XIX

SOIL ANALYSIS

INTRODUCTORY *

The general purpose of soil analysis is to give a quantitative expression of the constitution and properties of the soil. It is complementary to the description of the morphology of soil in the field in that it gives greater precision to the specification of certain properties such as texture and reaction; it is also supplementary in that it reveals and expresses certain properties, such as the composition of the clay fraction that cannot be estimated in a field examination.

Soil analysis is essentially the laboratory examination of soils as distinct from the field examination. It is true, of course, that soil analysis has been carried into the field by certain methods of rapid examination, yet such methods are generally only rough improvisations, except in those cases where, as for example in soil moisture studies, the essence of the method consists in its being carried out under field conditions.

Considering soil analysis as ordinarily practised, we may distinguish two distinct aims. Firstly, there is the practical aim of discovering and assessing plant nutrient deficiencies with a view to making recommendations on the manuring of crops. Under this head we may include the examination of the soil to discover the presence of constituents harmful to plant growth. To the practical man this is what is usually understood by soil analysis, and the num-

*The purpose of the present chapter is to discuss the significance of soil analysis. For details of actual methods the reader is referred to C. H. Wright, Soil Analysis, Murby and Co., London.
ber of analyses made annually in Great Britain for advisory purposes runs into tens of thousands. Soil analysis, or as it is more usually called, soil testing, is thus an important factor in British agriculture and, indeed, of the agriculture of all progressive countries. Just as in a modern industrial plant there is laboratory control at each stage, so in a modern farm manurial practice should, ideally, be based on the regular testing of each field.

There is, however, another purpose in soil analysis, namely, the exact and quantitative specification of a soil for the purpose of ascertaining its constitution and of comparing it with other soils. This purpose is primarily scientific and underlies all attempts to understand the genesis of soils and to arrange them in an ordered classification. The data obtained by this type of soil analysis are above all necessary to supplement descriptions of soil profiles in the field, which otherwise run the danger of becoming too qualitative and subjective. It is with this aspect of soil analysis that we shall be mainly concerned in the present chapter.

In considering methods of soil analysis, we may distinguish between absolute and conventional methods. In an absolute method, a determination is made of the total amount of a given constituent present in the soil, or of the amount falling in a definite category of the soil. The determination of the total nitrogen content of the soil is an absolute determination, as is also the determination of the ammoniacal nitrogen content. Similarly we may characterize the determinations of exchangeable bases as absolute determinations. Wherever it is possible to distinguish categories of constituents in the soil, the natural corollary is an absolute method of determination.

On the other hand, we have conventional methods. Many of the methods used in advisory work are of this type. They are generally the outcome of practical requirements and are of slight significance apart from their practical utility. In the determination of "available" phosphoric acid by the citric acid extraction method, it is necessary to
prescribe every detail of working; for any variation, for example, in the ratio of solvent to soil, or in the time of extraction, may lead to different figures being obtained. The phosphoric acid thus determined does not represent any category of soil phosphorus that can be defined in terms other than those of the method by which it is obtained. Methods such as these can have only a temporary value and must be superseded when fuller knowledge of the constitution of the soil enables us to distinguish and determine the different categories of soil constituents.

From the standpoint of the study of the soil as a pure science, the chief end of soil analysis is to characterize the material with which it deals. An ideal system of soil analysis should give a complete quantitative description that will serve as a basis for comparison and classification. The need for such a system of analysis is abundantly evident when we consider the qualitative terms in which many of the most important groups of soils have been described. The task of soil classification cannot be considered complete until these subjective descriptions have been translated into definite quantitative data. Such data are as important as the physical and chemical data necessary for the description of chemical compounds.

MECHANICAL COMPOSITION

The first information required about a soil is its mechanical composition. The practical man, confronted with a soil, will always wish to know whether it is a clay, a loam, or a sand, and the pedologist will seek the same information expressed in quantitative terms. Mechanical composition gives quantitative expression to a fundamental property of soil, unaffected by the ordinary vicissitudes of cultivation and cropping. The results of mechanical analysis can be most readily interpreted when shown in the form of a summation curve, for, as we have seen in Chapter II, different types of soil give characteristic curves. Thus, there are characteristic curves for highly weathered soils, for
mechanically weathered soils, for alluvial soils, and for soils of mixed origin.

Mechanical analysis is of particular value in controlling field estimates of texture. It should be noted, however, that mechanical composition is a property of the soil which, although closely correlated with texture, is quite distinct from it. Texture relates to the physical properties of the soil under cultivation and is essentially a property that obtrudes itself on the tiller of the soil. With soils having the same general mineralogical composition and organic matter status, mechanical composition reflects texture fairly closely, so that the concordance between field judgement and laboratory data may be very satisfactory. The character of the clay itself affects the texture; with highly sesquioxodic soils such as occur in the tropics, the tendency is to under-estimate the clay content.

The most important single figure in mechanical analysis is the clay content. If the clay fraction has been appropriately defined, it enables us to state what proportion of a given soil consists of the reactive colloidal products of chemical weathering, as distinct from the unweathered material, which is relatively inert and acts simply as a framework or skeleton for the reactive constituents.

IGNITION LOSS

The loss in weight suffered by an oven-dry soil on ignition was formerly used as a measure of the proportion of organic matter present. It was always realized, in the case of soils containing much clay or calcium carbonate, that the results were likely to be of little value, and even with lighter soils, small amounts of organic matter cannot be shown by ignition loss. Recently, however, this determination has acquired new importance, for B. A. Keen and P. R. H. Coutts\(^1\) have shown that it does afford an approximate indication of the colloidality of the soil and is highly correlated with the "sticky point" value. G. J. Bouyoucos\(^2\) has proposed a method for determining the combined water of soils by distillation. This procedure offers considerable promise
since by subtraction from the ignition loss of the figure thus obtained it is possible to make an estimate of organic matter.

CALCIUM CARBONATE

Calcium carbonate is usually determined and shown with the mechanical analysis of a soil. The presence of calcium carbonate, except where it occurs as comparatively coarse isolated fragments, indicates base-saturation and, in natural mature profiles, is characteristic of certain horizons of soils of the chernozem and chestnut earth group. Calcium carbonate may also occur, under certain circumstances, in horizons affected with ground-water. Among immature soils, the rendzinas are characterized by the presence of free calcium carbonate as, for example, in the chalk soils of South-East England and Northern France. In soils of the humid regions—the "pedalfers" of Marbut—calcium carbonate is only a fugitive constituent and does not occur in mature profiles.

ORGANIC MATTER

Of equal importance with mechanical composition is a knowledge of the organic matter content of the soil. The greater proportion of soil organic matter consists of colloidal material, which, with the clay, forms the colloidal complex. The organic matter is also the seat of micro-biological changes in the soil. It is obvious, therefore, that a knowledge of the amount of this constituent is of the highest value in characterizing a soil. Its direct determination has hitherto been a matter of some difficulty, although the method of Bouyoucos, to which reference has been made above, appears to offer possibilities. It is more usual to obtain a measure of it from the determination of organic carbon, using a conventional factor (1.724) to convert to organic matter. The organic carbon is determined by dry combustion or by wet combustion methods. In cultivated soils, it is sometimes necessary to make a correction for elementary carbon in the form of coal or cinder.
The determination of the organic matter content is only the first approach to a characterization of the organic portion of the soil. There are obvious differences between 5% of organic matter in a typical red loam, a black earth, and a podsol, respectively. It is, therefore, desirable to have methods for the fractionation of organic matter. Various methods have been proposed, such as the determination of humified organic matter by hydrogen peroxide, and the determination of pentosans, lignins, and celluloses; but, until the proposed methods have been investigated over a wider range of soils, the significance of the results yielded by them is not sufficiently evident for their adoption in routine analysis. It is probable, however, that investigations now in progress will eventually lead to the recognition of methods for the characterization of organic matter that will be of the highest value in soil classification.

NITROGEN

The determination of nitrogen is often made without any clear idea as to its significance. For purely practical purposes, it can rarely serve as a measure of the nitrogen status of the soil. As a measure of soil organic matter, it is highly untrustworthy because of the great variation in the nitrogen content of organic matter. In recent years, a considerable amount of work has been carried out on the carbon-nitrogen ratio of soils, and, whilst the significance of this ratio is not completely understood, some of the indications are sufficiently suggestive to render it highly desirable that it should be determined over as wide a range of soils as possible. It may yet prove a valuable aid to the definition of soil groups.

The significance of the ammoniacal nitrogen content of soils is not sufficiently evident to recommend its general determination. Except in acid organic soils, it is generally present in very small proportions. The nitrate nitrogen varies so greatly in amount in the same soil that it can scarcely be used as an aid to description.
TOTAL ANALYSIS

In a perusal of the literature of the subject, the student will find numerous descriptions of soils and soil profiles in which the complete analysis of the total soil is given. Such analyses, involving the determination of silica, sesquioxides, titanium oxide, alkalies, and alkaline earths, represent a very considerable mass of effort, yet the value of the information to be derived from them is, in the writer's opinion, of an entirely lower order of magnitude, since they fail to distinguish between the unweathered minerals and the weathering complex. They may be compared with results of analyses of entire animals made by the pioneers in animal nutrition. Total analysis may find some use in connexion with methods for estimating the degree of weathering of soils.

EXAMINATION OF THE CLAY FRACTION

So far as the mineral portion of the soil is concerned, the clay fraction is the seat of most of its chemical reactivity. Further, it is in most soils approximately identical with the weathering complex. For these reasons, its specification is of the highest importance. For many years it has been customary to express the composition of the clay by the molecular ratios $\text{SiO}_2 : \text{Al}_2\text{O}_3$, $\text{SiO}_2 : \text{R}_2\text{O}_3$, and $\text{Al}_2\text{O}_3 : \text{Fe}_2\text{O}_3$. Whilst the information given by these ratios sheds considerable light on the nature of the clay and on the trend of profile development, it has long been realized that the information thus given is far from adequate to specify the clay constituent. During recent years, considerable progress has been made in X-ray analysis and it is now possible to indicate the approximate proportions of the different types of clay minerals present.

Valuable as this information is, however, the specification of the clay complex is incomplete unless it includes also the necessary data about the non-crystalline constituents, in particular iron and aluminium oxides and free silicic acid.
It is this kind of information that is given by the method of O. Tamm\(^8\). Other methods are also available, including those of M. Drosdoff and E. Truong\(^4\) and of H. G. Dion\(^5\) for determining free iron. The value of these methods is that they give an insight into the character of the mobile or active colloidal material in the soil and may therefore provide a more sensitive index of eluviation and illuviation than total clay analyses or mineralogical data. The development and use of these methods can, therefore, be strongly recommended.

**EXCHANGEABLE BASES**

The properties of soils are, as we have already seen, so strongly affected by the nature and amount of the exchangeable bases present that any description of soil must include data for these constituents. For most soils of the humid regions, about 80% of the reaction value of the total bases, exclusive of hydrogen, is represented by exchangeable calcium, and therefore it is generally sufficient to determine only this base. In saline and alkaline soils, however, exchangeable sodium forms a considerable if not principal proportion of the exchangeable bases, with or without free sodium salts. The determination of exchangeable bases in soils containing notable amounts of free calcium carbonate is generally unnecessary, except in saline and alkaline soils, but may be carried out on typical samples.

In most soils, calcium is the dominant exchangeable base, and a determination of exchangeable calcium gives a fair measure of the total exchangeable bases present. For more accurate work, it is often desirable to determine the other bases and also the reaction value of the sum of the exchangeable bases. Where this value is reached by summation of separate determinations, it is usually somewhat magnified by the inclusion of soluble salts, principally calcium sulphate and chloride. This is avoided in the acetic acid method of R. Williams\(^7\).
The reaction of soils, generally expressed as pH, the negative logarithm of the hydrogen-ion concentration, has been the subject of a large amount of investigation. Before discussing its significance, it is necessary to direct attention to the somewhat conventional and arbitrary character of the results obtained. When we speak of the pH of a soil, we cannot achieve a precise significance unless the conditions of determination are exactly stated. The pH of a dry soil would be meaningless, since active hydrogen-ions can only exist in solution. Determinations of pH therefore refer to soil-water pastes or suspensions and, since different results can be obtained according to the soil-water ratio used, it is necessary to specify the actual ratio. It has been recommended by the International Society of Soil Science that a soil-water ratio of $1:2.5$ should be used, but it should be clearly realized that, even where this recommendation is followed, the results are only of comparative value and are no absolute measure of a soil property.

It should be noted that the reaction of a suspension of a soil, even under standardized conditions of determination, is governed not only by the degree of saturation of the complex acids of the colloidal complex, but also by the nature of the complex itself. Broadly speaking, the greater the ratio of the acidoid groups, silicic acid and humic acid, to the basoid groups, aluminium oxide and ferric oxide, the more acid will be the reaction at a given degree of saturation. A lateritic soil may contain very small amounts of exchangeable bases and yet show a pH approaching neutrality, whilst a soil rich in organic matter or with a highly siliceous clay complex might show a pH of 4 or even less at the same degree of saturation. Low pH values may also be due to the presence of mineral acids.

In spite of these limitations, pH data do give an indication of the relative base-status of similar soils and are of significance in connexion with many of the most important pedogenic processes. The lowest figures are found in
peats and humus podsols, whilst the highest figures are found in alkaline soils. Red podsolic soils, red loams, brown forest soils, and black earths, occupy intermediate positions.

CONSTITUENTS SOLUBLE IN STRONG ACIDS

Proceeding from the assumption that regulated extraction with strong mineral acids can separate the weathering complex of the soil from the unweathered portion, many investigators determine the amounts of silica, alumina, ferric oxide, and other bases soluble in hydrochloric or sulphuric acid under conventional conditions of working. In the writer's opinion, however, there is no evidence that any acid extraction can succeed in dissolving the weathering complex without attacking unweathered minerals in addition. Nevertheless, the results obtained by such methods may afford indications which may be of practical value as to the reserves of plant nutrients in the soil. Phosphoric acid is frequently determined in acid extracts, but no evidence is yet available that the amount thus brought into solution represents any definable category of the phosphorus compounds of the soil.

SOLUBLE SALTS

In certain classes of soils, it is desirable to make determinations of the content of soluble salts. As a complete analysis of the soluble salts is generally a tedious operation, it is often considered sufficient to obtain an indirect estimate by determining the electrical conductivity of aqueous suspensions.

PHYSICAL DETERMINATIONS

An account of soil analysis would be incomplete without some reference to the use of physical determinations for the description and definition of soils. But, apart from mechanical analysis, which is essentially a physical determination, it must be admitted that methods are still in process of evolution, and that no generally accepted system of physical determinations exists. It may be helpful, however,
to review briefly those physical properties of the soil which, from the standpoint of soil definition, call for quantitative expression.

PORE-SPACE AND VOLUME WEIGHT

A knowledge of the variations in pore-space and volume weight throughout the soil profile would doubtless be of considerable aid to the definition of the principal soil groups. But whilst the determination of pore-space and weight per unit volume offer no difficulty with laboratory samples, determinations under field conditions are laborious and, except in almost stoneless soils, beset with considerable experimental difficulties. In spite of the importance of the information sought, it must be admitted that the determination of pore-space has not yet become one of the routine investigations of the soil worker.

SINGLE-VALUE DETERMINATIONS

In recent years, attempts have been made to express those properties of the soil associated with their colloidal character by means of so-called "single-value" figures. Of those which may more properly be called physical determinations, hygroscopicity, moisture content at the "sticky point," and heat of wetting, have been most widely used.

Whilst the earlier workers on hygroscopicity attempted to determine full hygroscopicity in equilibrium with a saturated atmosphere, recent workers use atmospheres below saturation point. There is, at present, no uniformity of procedure.

The determination of the moisture content at the "sticky point" has, in recent years, attained considerable vogue among British workers, and since it does appear to have a definite physical meaning, i.e., the moisture content at which the soil gels are completely saturated, it should prove of great value in defining and describing soils. As it is determined with comparatively little labour and expense, it may be recommended for routine workers and will doubt-
less prove of increasing value as results are accumulated and examined in relation to other soil properties.

Heat of wetting has been used by certain American and continental workers, but has not come into general routine practice.

COLOUR DETERMINATION

The colour of soils is an important aid in the recognition of the soil groups. Such groups as laterites, terra rossa, podsol, and black earths, have colours which are highly characteristic. These are almost invariably described in qualitative terms, with the result that, whilst a worker familiar with podsol has a clear idea of the range of colours met with in these soils, his descriptions will not convey an accurate impression to a worker familiar with tropical soils who may, indeed, use similar terms to describe unmistakably different colours.

The American soil survey workers have made efforts to define soil colours by methods such as the Munsell colour scale. Unfortunately, standards for comparison are neither convenient nor inexpensive, and no method of colour description has come into general use. The need for a routine method is very urgent, and it may be hoped that it will be met in the near future. (See p. 261.) In the meantime, the writer has found the Ridgway colour atlas very convenient for laboratory determinations of colour.

SOIL ANALYSIS FOR ADVISORY PURPOSES

If our knowledge of the constitution of the soil were more complete, there would be no need to draw a distinction between soil analysis for scientific purposes and soil analysis for advisory purposes, because a complete specification of the soil would not only describe the soil but also supply all the information necessary for giving advice to farmers on the manurial treatment of their land and for diagnosing causes of poor performance. In the absence of any complete under-
standing of the constitution of the soil and of the chemical and microbiological changes taking place in it, there has grown up in each country a body of analytical methods, mainly empirical, designed to answer specific questions. For example, there is a wide choice of methods that purport to answer the question: Is this soil in need of a dressing of lime or chalk and, if so, at what rate should it be applied? To answer this question it is necessary (1) to make certain analytical determinations and (2) to interpret the results of these determinations. The very wide choice of methods reflects the great divergence of opinion about the relevant determinations, whilst the interpretation of the analytical data requires that they shall be related to a wide range of circumstances such as crop, season, and possibly economic conditions, in the light of available experience. It is therefore incorrect to speak of determining the "lime requirement" of a soil. The actual requirement is largely a matter of opinion in the light of certain analytical data.

Soil analysis for advisory purposes as carried out is largely empirical, yet, as practised by experienced advisers, can give valuable aid to the practical man. The underlying principle of much of this work is the determination of so-called "available" plant nutrients. It was realized at an early stage that total analyses of the soil can give little useful information, and many methods of extraction have been proposed whereby the available may be distinguished from the unavailable plant nutrients. Thus, B. Dyer originated the widely used method of extraction with 1% citric acid. Other methods involve the use of such extractants as N/5 nitric acid, N/200 sulphuric acid, and N/2 acetic acid.

It is not our intention to discuss these methods in detail but it may be helpful to examine some of their underlying principles.

In the first place it should be realized that there is no clear distinction between "available" and "unavailable." For example, there are many categories of soil phosphorus, both inorganic and organic. Some are more and some are
less available; yet even the most insoluble may contribute to some extent to the phosphate nutrition of plants.

Secondly, whatever extractant is used, each category of a given plant nutrient will have a certain solubility in it, although some will be more and some less soluble.

Thirdly, in most methods of extraction the amount of a particular ingredient extracted will be affected by such factors as ratio of soil to solvent, time and method of shaking, and temperature. With few exceptions, the methods are conventional.

Fourthly, whatever method is used it is necessary to interpret the results. To answer the simple question as to what constitutes deficiency, laboratory results must be related to field behaviour, and this implies the setting up of carefully designed field experiments with different crops over a number of seasons.

The choice of suitable analytical methods and their correct interpretation are the foundations on which soil analysis for advisory purposes rests. There are great difficulties at each stage. An analytical method must not only give significant results; it must also be adaptable for use with large numbers of samples. Here spectrographic and photo-electric methods have proved of great value. The interpretation of analytical data presents its own special difficulties and is often highly subjective.

Given the difficulties of using soil analysis as a basis for advisory work, it is not surprising that other methods have been sought for diagnosing the plant nutrient status of soils. Of these alternative methods, the pot technique of E. A. Mitscherlich, in which the response of crops grown in a given soil to increasing doses of fertilizer is studied, has attained a wide use in Germany. Another approach is that by H. Neubauer and W. Schneider, in which seedlings grown on the soil are analysed for manurial ingredients. There is also the Aspergillus method of H. Niklas, G. Wilsmeier, and F. Kohl, in which fertilizer requirements are deduced from the growth of the mould Aspergillus niger.
on preparations of the soil with varying additions of nutrient elements. The concordance of the results of these methods with soil analytical methods has been discussed by E. A. Mitscherlich\textsuperscript{11}. It seems clear that no single method will give trustworthy information in all cases, but that there are a number of methods that can prove satisfactory in a majority of cases.

An entirely different attack on the problem is by using the growing plant for diagnostic purposes. Pioneer work in this field was done by H. Lagatu\textsuperscript{12}, and the technique has been considerably developed during recent years by T. Wallace\textsuperscript{13} and his collaborators. Considerable use may be made of the diagnosis of nutrient deficiencies from discolorations and other appearances on leaves. Atlases have been prepared showing the symptoms of manganese, potassium, boron, iron, and other deficiencies on the leaves of different crop plants. This method has been reinforced by the development of sensitive colour reactions for assessing the status of leaf tissues in respect of some important plant nutrients. The method is probably capable of development and extension, but even when perfected must suffer from the disadvantage that the deficiency is discovered only after the plant has started into growth. For this reason, plant nutrient diagnosis, however satisfactory it may be in revealing a nutritional deficiency or imbalance, can never be so helpful to the practical man as methods that will enable him to take the necessary measures before crop growth begins. Diagnoses from foliar examination is probably more helpful in the case of tree and bush growth than in the case of annual field crops, where the deficiency may be discerned too late for remedial measures to be taken. Yet even here, early diagnosis may often permit action to be taken in time to save a crop. For example, manganese deficiency in oats can be diagnosed in the early stages and the immediate application of manganese sulphate in spray form can save the crop.

From what has been said it is clear that there is no generally accepted system of soil analysis in examination for
advisory purposes. Nor is such a system in sight, because given the importance of interpretation the widely differing conditions in different countries and regions must lead to a variety in advisory methods.

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CHAPTER XX

SOILS, PLANT GROWTH, AND AGRICULTURE

The purpose of this work is mainly to present an account of the soil as a natural object of study. We have purposely refrained from presenting the problems of the soil from the standpoint of crop production, partly on account of the complexity of the considerations involved and partly because a satisfactory understanding of the nature of the contract between the soil and the living plant must be preceded by a thorough understanding of the constitution of the soil. Yet an account of the soil as a natural object would be incomplete without some reference to its relationship to the higher plants which inhabit its surface. For this relationship not only constitutes the chief human interest in the soil, but is also of direct importance in the study of the development of the soil itself.

NUTRITION OF PLANTS

The dry matter of plants consists of organic matter, and mineral or ash constituents. The principal constituents of organic matter are:

- Carbohydrates, including sugars, pectins, and starches.
- Fibres.
- Oils, fats, and waxes.
- Proteins.

The carbon, which forms about 55-60% of the dry organic matter, is obtained by the process of carbon assimilation or photosynthesis from the carbon dioxide which forms about 0.03% of the atmosphere. This change consists
essentially in the formation of starch from carbon dioxide and water, probably through formaldehyde and dextrose as intermediate stages, with concomitant liberation of oxygen. The reaction, or chain of reactions, is endothermic and the source of energy is the sunlight. A further condition for the process is the presence of chlorophyll, the green colouring matter of plants. Except in so far as carbon dioxide is produced in the soil by the oxidation of organic matter, the soil cannot be considered as of direct importance in carbon assimilation. The hydrogen and oxygen of organic matter are obtained ultimately from the soil moisture.

The remaining elements of the dry matter of plants, namely, nitrogen, and the elements present in the ash, namely, sulphur, phosphorus, calcium, magnesium, potassium, iron, silicon, and all the other elements which, though not in all cases essential to growth and reproduction, normally occur in plants, are obtained from the soil.

Additions are constantly being made to the list of elements considered necessary for plant growth. Thus, during recent years, manganese, boron, copper, zinc, and iodine have been shown to be essential. They were overlooked by earlier workers on account of the minute quantities required relative to the other nutrient elements. The traces of impurities present in the materials used for culture solutions were often actually sufficient to supply these elements. It is likely that the list of nutrient elements will be further enlarged. Progress is limited by the inadequacy of ordinary analytical methods. The adoption of spectrographic methods has enabled considerable advances to be made in the survey of the essential elements for plant growth and their occurrence in soils. In the meantime, it is possible that some of the obscurer cases of infertility in soils may be due to unascertained deficiencies in trace elements.

It is generally supposed that the elements thus obtained, often termed plant nutrients, enter the plant root by absorption from the soil solution; and much of the theory of plant nutrition has been built up on the basis of experiments in
which plants are grown in so-called culture solutions. It is possible, however, that the mode of absorption by the plant may not be so simple as this theory implies. N. M. Comber has put forward the suggestion that the colloidal material of the soil and the colloidal material of the plant roots may form a single system within which translocation of material may take place. Whatever be the nature of the contract between soil and plant, it is evident that a principal duty of the soil is to furnish the growing plant with an adequate and balanced supply of the nutrient elements. In many soils, mycorhizal associations may be of great importance for plant nutrition.

There are, however, other factors involved in plant growth. Apart from the water required for the actual material of plants, water is required by the roots of plants to make good the constant losses by transpiration from the leaves. It has been found that, for every part by weight of dry matter formed, from 200 to 1,000 parts by weight of water are required. The supply of water to plants is intimately bound up with the supply of oxygen needed for the respiration of the root system of plants. An excess of water, whilst not necessarily disadvantageous in itself, restricts the air supply to roots. On the other hand, an excess of air, not in itself harmful, implies a deficiency of moisture. Between the two extremes lies the optimum, which is often considered to be fulfilled by a 50% saturation of the pore-space of the soil with water.

The ease with which the soil yields up water to plants diminishes as the moisture content decreases below the optimum. Finally, a point is reached at which the water is so firmly held by the soil colloids that the losses by transpiration can no longer be met, and wilting, i.e., loss of turgidity, occurs. The moisture content at the wilting point is roughly represented by 4.2 on the pF scale (p. 283). It is thus evident that the moisture present in a soil is not entirely available to plants, but is subject to a deduction that is dependent on the colloidality of the soil.
Each plant has its own minimum, optimum, and maximum temperature for germination and also for adult growth. The value of any particular situation or soil for plant growth is therefore relative to the plants which can grow there. Plant growth cannot take place below 0°C, but certain arctic and alpine species can grow at temperatures immediately above freezing point. At the other extreme are plants which can tolerate temperatures up to 45-50°C. The optimum temperature for growth varies considerably among different plants.

Generally speaking, the temperature of the soil is closely related to that of the air; but, as we have seen in Chapter IV, minor differences may occur among soils even under the same climate. In regions where the growing season is limited by winter, such minor differences may be of considerable importance in their effect on the beginning and duration of the year’s growth.

The next factor to be considered is negative, namely, the absence of injurious or toxic substances. The most obvious cases of inhibition of plant growth due to this group of factors are seen in soils in the vicinity of lead, copper, tin, or zinc mines, where the presence of compounds of these metals partially or entirely excludes vegetation. The considerations advanced in discussing "trace" elements in nutrition are applicable to injurious elements. Cases of infertility may occur where the injury is due to small traces of elements whose presence in the soil has hitherto been unsuspected.

Another type of injury due to harmful constituents is that caused by the presence of excess of soluble salts, most commonly sodium chloride, sulphate, or carbonate. There is a considerable variation in the tolerance of different plants to soluble sodium salts. Indeed, some plants, natives of the sea shore and of saline soils, may be described as halophytic. Most economic plants, however, can tolerate only small concentrations of salts in the soil moisture, and the presence of such salts is a serious drawback to the agricultural use of soils in which they occur.
Finally, there is the possibility of injury or limitation of plant growth through the presence of acids, or, defining the case more strictly, the prevalence of an excessive concentration of hydrogen-ions in the soil moisture.

The problem of soil acidity in its relationship to plant growth has been for many years the subject of laborious investigation and active debate. That certain soils fall below satisfactory productiveness by reason of the prevalence of conditions that can be remedied by the application of dressings of lime or calcium carbonate cannot be doubted. That acidity in itself is the injurious factor cannot be so generally maintained. It is certainly true that an extreme degree of acidity, such as is rarely found in ordinary soils, is injurious to plant growth. It is also true that a reaction slightly on the acid side of neutrality may inhibit the activity of the nitrogen-fixing bacterium Azotobacter, and in so far as this activity is a necessary part of the nitrogen economy of the soil, acidity is harmful. Otherwise, most plants can tolerate a fairly wide range of reaction, although the optimum reaction for individual species varies somewhat.

The problem is complicated by the fact that soil acidity cannot enter into the problem of soil fertility as an independent factor, for it is correlated with a deficiency of available calcium and also, in some cases, with the presence of actively toxic aluminium compounds. Indeed, M. Trénéel and F. Alten have demonstrated in culture solutions that aluminium, active below pH 5.0, acts as a specific root poison. Further, in some cases, soil acidity may be correlated with undesirable physical conditions. It is thus not easy to assert generally that soil acidity, per se, is a limiting factor. The limitation may actually be due to lack of available calcium, the presence of active aluminium, or the prevalence of an undesirable physical condition.

**SOIL FERTILITY**

We have enumerated the principal groups of factors affecting the growth of plants: the list may not be exhaus-
tive, for other factors may be involved. For optimum growth all factors must operate satisfactorily. The unsatisfactory operation of a single factor, such as the supply of moisture or available nitrogen, may offset the satisfactory operation of all other factors. Where it is possible to use a quantitative expression, e.g., nitrogen supply, a relationship can be traced between the magnitude of the factor and the yield obtained. Many equations have been devised to exhibit this relationship. The simplest is that which assumes that the increment in yield for each successive unit increment in the factor is proportional to the deficit from the maximum yield obtainable. In the form of an equation this may be stated as—

$$\frac{dy}{dk} = k(A - y).$$

Where $y$ is the yield, $x$ is the growth factor, e.g., nitrogen
supply, \( A \) is the maximum yield and \( k \) is a constant. If we term one unit the amount of the growth factor necessary for 50% of maximum yield, the relationship between the magnitude of the growth factor and yield would be as follows:

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<tr>
<th>Units</th>
<th>Yield as % of maximum</th>
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<tr>
<td>1</td>
<td>50</td>
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<tr>
<td>2</td>
<td>75</td>
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<tr>
<td>3</td>
<td>87.5</td>
</tr>
<tr>
<td>4</td>
<td>93.7</td>
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<tr>
<td>5</td>
<td>96.7 and so on.</td>
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These relationships are illustrated by the curves in Fig. 22. In curve A is shown the theoretical response to successive unit additions of a plant nutrient, for example, nitrogen, assuming that all other factors are operating satisfactorily. Curve B shows the way in which the yield may be modified when one or more of these other factors operate unfavourably. Here the maximum yield obtainable by increasing the supply of the nutrient in question is much smaller, but the form of the response curve is similar.

Curves showing the relationship between yield and the magnitude of any growth factor such as nitrogen or phosphorus supply are known as \( \text{response curves} \). The response of a crop to a given addition of any plant nutrient will depend on two circumstances, namely, (1) the amount of the nutrient already present and (2) the maximum yield possible as determined by the operation of the remaining factors. Thus a low response to a dressing of fertilizers may be obtained either because the nutrient is already present in abundant supply or because some other factor or factors, e.g., water supply or lime supply, are so deficient that only a small yield is possible whilst these deficiencies are unremedied.

The actual yield of a particular crop will depend on a great variety of factors, including seasonal variations and even casual factors such as the incidence of diseases and pests, so that it is somewhat difficult to assign to any soil a
numerical assessment of productivity. If such an assessment were possible it would be necessary to take into account (1) the soil factors, (2) the site factors, including climate, and (3) type of management and crops or stock raised.

The concept of soil fertility is, as we have seen, very complex. Apart from this, it is necessary to note the different sense in which the term is actually used. In different contexts the term fertility may bear any of the following senses:

(1) Soil fertility may be taken to mean the productivity of the soil, as measured by the yield per acre. This is sometimes termed "productivity rating" and is arrived at by calculating a weighted average of the principal crops grown each yield being expressed as a percentage of a supposed maximum. This estimate may be further elaborated by giving the rating for ordinary management and for the best type of management.

(2) Soil fertility may be regarded as a measure of the response given by the soil to expenditure in labour and materials. It is this that principally determines the agricultural value of a soil and the rent that the farmer is prepared to pay for it, although other factors such as accessibility to markets and other economic factors may also play a part in determining this value.

There is no absolute scale of fertility, because the use to which the land is put must be taken into consideration. Some soils are chiefly valuable as pasture, and the most famous fattening pastures retain their reputation and value only when used as pasture. Other soils may be valued highly for arable farming, others for market gardening or fruit culture. In the case of irrigated lands the natural productivity may be almost negligible, but the return for cultivation when irrigation works are installed may be very great indeed.

(3) A third sense in which the term fertility is used is that of plant nutrient status. This rather restricted use of the term is more common in the United States than in this
country. Yet many writers may inadvertently use the term in this sense, without explanation.

It would help considerably towards an adequate treatment of the problems of crop growth if terms such as soil fertility and soil productivity were used in a more precise sense.

PROBLEMS OF CROP GROWTH

We may review briefly the problems which the pedologist is called upon to solve when he assumes the rôle of agricultural chemist and attempts to give assistance to practical agriculture.

Firstly, he is called upon to show how soil productivity measured in yield per acre or hectare can be most economically increased. To answer these questions he must know the effect of cultivation and management, crop rotation, and manurial practice on plant growth under the particular conditions of climate and soil encountered in his district or country.

Secondly, given the particular type of land utilization, he should be able to lay down the conditions under which yields may be maintained at an economic level and the soil guarded against deterioration and loss.

Thirdly, he must be able to prescribe the measures to be taken, if economic conditions should demand a change in the system of land utilization.

It must be admitted that the present state of our knowledge falls far short of that necessary for the adequate solution of these problems—in so far as they apply to the actual growth of crops. Yet their solution is urgent at the present day, when there is such a serious world shortage of food. Plan and order are necessary alike for expansion and restriction.

There appear to be two approaches to the solution of the problems of soil fertility and soil management. In one, knowledge is acquired by observing the effect on crop production of various methods of cultivation, management,
cropping, and manuring. By isolating the variables and applying the latest statistical methods the results thus obtained can be given the highest significance. The results of such experiments contribute little to increasing our knowledge of the master-problem of the contract between soil and plant, yet they are necessary in order to solve immediate problems.

There is also another approach, namely, the study of the master-problem of plant nutrition and growth. The pursuit of fundamental knowledge needs no justification, for this is the classical road by which science has passed on its gifts to practical life.

Field experiments must always be carried out, particularly in undeveloped countries face to face with new problems, but the writer would deplore any avoidable diversion of material resources and creative thought into this pedestrian type of investigation, which opens out no new horizons, but merely answers the commonplace questions it is required to solve.

If it is conceded that the surest approach to the practical problems of crop production is through fundamental research on plant physiology, the task of the pedologist is to elucidate the constitution and properties of the soil in order that the plant physiologist, when he carries his investigations from the artificial culture solution to the soil itself, may find a medium whose reactions are accurately known and whose significant properties can be specified in precise quantitative terms. The progress which has been made in the study of the soil itself during recent years encourages the belief that the time is ripe for an attack on the problems of plant nutrition along the lines indicated in the above discussion.

SOILS AND AGRICULTURE.

The earliest cultivations of soil were in arid and semiarid regions, and it has been shown that in the westward migrations into Europe the first settlements were in forest-free regions, the loess areas, which in their general character
resemble steppe. For primitive agriculture, the steppe or prairie offers a much better prospect for successful crop production than the forest soils of humid regions. Under steppe, the climatic conditions, though drier than in humid regions, are more dependable. There is generally sufficient moisture after the winter for the establishment of crops, whilst the hot dry summer is favourable to ripening and harvest. Further, the moderate leaching tends to the conservation of plant nutrients and the inhibition of the deteriorative processes which operate in humid soils. The chief problem is water supply and yields are determined mainly by this factor. It is significant that the principal wheat areas are on the black earths and the chestnut earths.

All agricultural soils were formerly virgin soils. In long-settled countries most of the soil has been in agricultural use for centuries, and the profile characters of the primitive soils have been profoundly altered during the centuries. In newer countries, such as the United States and the Argentine, many soils still retain some of the characteristics of the virgin soils from which they have been reclaimed.

A comparison of virgin with cultivated soils is instructive. Under virgin conditions, soils and vegetation form a system that is in a state of dynamic equilibrium. The mineral plant nutrients abstracted from the soil in each season’s growth are returned in leaf fall or by the decay of herbaceous and annual plants. It is true that in forests there is an annual storage of plant food due to tree growth, but in the absence of human interference this material is ultimately returned to the soil as trees die and decay. The only losses are in the form of drainage. Against this may be offset the gains from the weathering of minerals in the soil. Under such conditions it is not surprising that luxuriant vegetative growth is possible with a comparatively small supply of available plant food, of which a considerable proportion is in circulation between vegetation and soil.

This equilibrium is completely destroyed when the virgin soil is brought into cultivation. The available plant
food in circulation suffices to give a satisfactory or even a heavy crop at first, but as each crop is removed from the soil the supply shrinks and yields fall off. If cropping is continued, yields are stabilized at a low level determined by the balance between losses by drainage and crop removal on the one hand, and the gains due to weathering of minerals and nitrogen fixation on the other. If production is to be at a higher level the inevitable losses by drainage and crop removal must be made good by addition of manures and fertilizers. In the case of nitrogen, fixation by bacteria can balance to some extent, in some cases completely, the losses by drainage and removal of crops.

We may now proceed to consider the probable consequences of bringing into cultivation different kinds of virgin soil. For this purpose we shall consider four typical cases, namely:

1. Podsolic soil, associated with coniferous forest or heath.
2. Brown earth or grey-brown podsolic soil, associated with deciduous forest.
3. Prairie soil, associated with tall-grass vegetation.

1. Considering first the podsolic soil. In its natural state it is characterized by very low base-status, an organic layer of raw humus, and a loose structure in the mineral A-horizon. This soil, even if used under an "acid" type of agriculture for such crops as oats, potatoes, and hay, requires immediate dressings of lime or chalk to correct the lime deficiency, to decompose raw-humus, and to promote the development of crumb-structure. It would certainly need phosphatic and potash fertilization, and would probably need nitrogen also to induce biological activity. In permanent cultivation podsolic soils would need regular calcareous dressings in order to make good the inevitable losses by leaching.
The humid climate would favour utilization as grassland, but grass on such soils is liable to degeneration unless carefully managed. It is therefore advisable, after a number of years in grass, to introduce an arable break and to seed down afresh. This is the type of agriculture practised in Wales. With attention to calcareous and phosphatic dressings a fairly high level of productivity is possible; otherwise degeneration is rapid.

(2) Brown earths have an advantage over podsols in their better base-status, owing either to the character of the parent material or to the drier soil climate. Raw humus is absent and the soil structure is better than in the first case. The drier climate is more favourable to arable husbandry and grain crops would offer fewer anxieties than under the wetter climate of the podsols. Although good crops would be obtainable at first, the need for calcareous, phosphatic, and potash dressings would soon be felt. Under continuous arable cultivation there would be the danger of organic matter being lost by biological oxidation. This would be accompanied by loss of structure and the danger of erosion in climates where periods of intense rainfall occur. Although the climate may not be humid enough for exclusively grass husbandry, temporary leys are desirable to maintain the organic matter status. For the same purpose, animal husbandry, with its production of farmyard manure, is of great importance.

(3) Prairie soils developed under tall-grass vegetation are characterized by higher base-status than the brown earths. They are also richer in organic matter and possess well-developed crumb structure. When first brought into cultivation they yield heavy crops, which are fairly well maintained for some years, even under continuous cultivation, particularly if rotations are suitable and if live-stock husbandry is practised. The initial base-status is higher than that of the brown earths and podsols but applications of lime may eventually become necessary. With the drier climate, the chief peril is depletion of organic matter. The
climate is favourable for cereals, whilst the maintainance of artificial grassland is rendered difficult by periodic droughts. Soils of this type occur in the most productive agricultural regions of the United States, as in Iowa, Illinois, and Missouri.

(4) Chernozem soils developed under short-grass vegetation have still higher base-status and better developed crumb structure. Good crops are obtained at once and are well maintained without the need for dressings of lime. The dry climate favours cereal production, but there is also the risk of occasional crop failure through drought. Under continuous and exhaustive exploitation, they lose organic matter with consequent deterioration in structure. They then become subject to wind erosion. Most of the serious wind erosion in the United States and Canada during recent years has occurred on these soils.

Each soil has its own inherent tendencies to deterioration; on the soils of humid climate, owing to lowering of the base-status, and in arid climates owing to loss of organic matter. The maintenance of permanent agriculture depends on the understanding of these tendencies, and the practice of types of agriculture that will counteract their operation. The serious consequences of neglecting to establish permanent systems of agriculture, and of robbing the accumulated fertility of virgin soils may be seen in the vast areas of ruined soils in the United States and in many parts of the tropics.

So far as Europe is concerned, with the exception of Spain and parts of Southern Europe, the utilization of the soil has proceeded along rational lines. In the opinion of the author, however, it is a mistake to rely too much on arable culture with the maintenance of fertility by the use of artificial fertilizers. A type of husbandry, such as that practised in Wales, in which the soil is periodically laid down in grass, may offer possibilities of economically increasing production. Land in grass accumulates fertility whilst providing food for grazing animals. When ploughed
up it has many of the characters of a prairie soil. It is, of course, difficult to establish temporary grassland in regions with hot dry summers, but considerable progress has been made in the technique of sowing down to grass. The introduction into the rotation of a period under grass might be of considerable value in the systems of agriculture practised in central and eastern Europe.

In spite of the more intensive agriculture of northwestern Europe, the cultivator must constantly struggle against conditions that militate against arable culture. One of the principal differences between soils of the chernozem group and soils of the podsolic group is in their structure. Whilst the calcium-saturated soils of the former class naturally tend towards the desirable crumb structure and demand a minimum of cultivation, the podsolic soils tend to assume the single-grain structure. The cultivator is therefore always attempting to establish a structure which is purely artificial. The leaching consequent on the high rainfall results in an impoverishment in mineral constituents, principally lime, which, if unchecked by applications of calcareous dressings, eventually results in deterioration, accompanied by loss of the desirable crumb structure. The actual labour of husbandry is increased by the fight against weeds and the anxieties of harvest operations. High yields can be obtained in arable farming in the humid districts of western Britain; but it should be realized that, since the natural vegetation is forest or heath, constant effort is necessary to prevent deterioration. It is perhaps no exaggeration to say that the efforts of the farmer in the humid regions are directed towards the production and maintenance of artificial chernozems. Having said this, it must also be said that some of the most successful agriculture and the most productive land occurs on these soils where the skill of the farmer is pitted against the tendencies of the soil to become impoverished.

The agriculture of a particular locality is the reflection of a complex of factors, of which climate, soil, and situation are the most important, since they impose certain limits on
the choice of crops for cultivation. But economic, social, and even personal factors also intervene, with the result that the agriculture of a country only partially reflects the variations in soil and climate within its borders, and may undergo changes from generation to generation in response to causes other than those arising from fundamental natural factors. The inter-relationships of these factors may be thus represented schematically:

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<tr>
<th>Geology</th>
<th>Climate</th>
<th>Topography</th>
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<tr>
<td>Natural soil properties</td>
<td>Natural conditions for plant growth</td>
<td>Cultivation and management</td>
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SOIL CONSERVATION AND PERMANENT AGRICULTURE

In a period of agricultural depression, when the tendency is towards restriction of effort, deterioration is likely to ensue, and to be most pronounced in those regions where conditions are most artificial. The consequences of "low farming" are likely to be less serious where conditions approach more closely to those of the steppe and the prairie.

The tendency to degeneration of soils in humid temperate regions is always present, but the highly developed agriculture of countries such as England, Holland, and Belgium, is sufficient proof that the struggle against it is no impossible task and that, with cheap fertilizers and skilful husbandry, crop yields can be obtained that would not be possible in drier regions.

Under the conditions of the humid tropics, the maintenance of an artificial system of land utilization is beset with even greater difficulties. The tendency to deterioration of virgin land brought into cultivation is far more rapid than in temperate climates. The high temperatures lead to a very rapid destruction of reserves of organic matter with conse-
quent deterioration in physical condition, whilst the torrential tropical rains, acting on such soils, produce vigorous erosion. It is not surprising, therefore, that there are many instances of newly broken soils becoming completely derelict within a few years through injudicious management.

Cultivated soils are, to a large extent, artificial and show profiles that may differ markedly from the corresponding virgin profile. From the moment that the original cover of natural vegetation is destroyed, the processes that have produced the original succession of horizons are modified, and new factors, leading to a new profile succession, are introduced. Loss of organic matter is the commonest consequence of cultivation and may be demonstrated under practically all climates when pioneer cultivation is instituted. F. T. Shutt gives an instance of a Manitoba prairie soil which showed a decrease of nitrogen content from 0.651% to 0.506%, and of ignition loss from 19.43% to 14.79%, as the result of 25 years of cultivation. The writer found a cultivated soil in Caernarvonshire to contain 0.490% nitrogen, and to have an ignition loss of 16.6%, whilst the adjoining waste, from which it had been reclaimed, contained 1.106% nitrogen and gave 43.1% ignition loss.

Loss of organic matter is not the invariable consequence of bringing land into cultivation. Where grass husbandry is practised, or where heavy dressings of organic manures are given, as in market gardens, the organic matter content of the soil may be maintained or even increased. In general, the loss of organic matter is least where livestock farming is practised, for it is then possible to return to the soil a considerable proportion of the crops raised and fed to stock, in the form of manure.

In many parts of the world, the most serious consequence of cultivation is erosion. We have already referred to the widespread erosion in the Eastern United States, which has taken place within comparatively recent times. Through uninstructed utilization of land in past generations, much irreparable damage has been done over large areas of
the tropics and sub-tropics. It is important, therefore, that future agricultural operations should be conducted with the fullest possible precautions against further losses.

The crops obtained immediately after bringing virgin land under cultivation are always bigger than succeeding crops. It is the common experience that yields fall off as the land is longer under cultivation. Under virgin conditions, the demands on plant nutrients are comparatively slight, for a large proportion of the nutrients extracted from the soil by plant roots is returned to the soil again in the form of plant residues. Under pioneer cultivation, the soil is cropped continuously and deterioration is inevitable. Firstly, there is a rapid exhaustion of plant nutrients as crop after crop is removed, and secondly, there is, in many cases, a physical deterioration in the soil consequent on loss of organic matter. This is manifested by a decreased water-holding capacity and an increased liability to erosion.

Even where the land is not brought under actual cultivation, uncontrolled grazing may lead to deterioration of the soil. Examples of this may be cited from South Africa, where injudicious grazing and close folding (kraaling) has led to a destruction of the natural cover. The consequence of this has been erosion by gullying, resulting in lowering of the water-table, and accentuation of the effects of drought.

The deterioration consequent on pioneer cultivation of virgin land may be checked where live stock husbandry is practised, for it is then possible to return to the soil a considerable proportion of the nutrients extracted from it by crops. Organic matter may also be maintained by green manuring. The introduction of artificial fertilizers, whereby the plant nutrient status of the soil can be controlled, marks a further stage in land utilization. With the present abundance and cheapness of fertilizers, the serious problems of continuous agriculture are mainly physical in character. Indeed, in some systems of agriculture, crops are grown almost entirely on added fertilizers, and the rôle of the soil is mainly physical. This is the case over a large part of the cotton lands of the United States.
The utilization of land for irrigation brings its own set of problems. In many instances, the continued addition of irrigation water has led to a rise in the water-table, whereby injurious salts have been brought to the surface. In other cases the irrigation waters used contain excess of sodium over calcium salts, with the result that the irrigated soil is gradually changed from a calcium to a sodium soil with consequent deterioration and ultimate disaster to crops. Fortunately, the behaviour of saline and alkaline soils has been so thoroughly studied that the pedologist can afford immediate help in proposing remedial measures for soils threatened by deterioration through alkalinization.

During the past 150 years the tendency has been towards an increase in the agricultural production of the world as a whole. This has been fostered by the settlement of virgin lands, by industrial development, which has rendered possible the application of machinery to agriculture, and by the astonishing progress of agricultural science, which has shown how greater yields may be extracted from the soil by the use of artificial fertilizers. Progress in the practice and science of agriculture goes on at an ever-increasing rate. It might appear, therefore, that the future food supply of the world is in no danger of proving inadequate to the needs of world population.

A more careful consideration of the position, however, may lead to a less optimistic view of the future position. In every continent there are vast areas that were once productive but are now represented by ruined soils of negligible agricultural value. This deterioration may be due to desertization, as in central Asia and North Africa, to wind erosion, as in the western states of the U.S.A., to water erosion, as in parts of Southern Europe, in the eastern states of the U.S.A., and in the humid tropics, or to the salinzation and alkalinization of irrigated soils, as in Mesopotamia, South Africa, and California. In every case, reckless exploitation of the soil has led to deterioration and ruin. In past centuries this has been less culpable because of the
limitation of man's knowledge of the soil and his ignorance of the probable consequences of the use that he was making of it.

When these limitations are added the disorganization due to wars and their aftermath, the prospect of abundant and cheap food throughout the world appears very remote, even although some countries may experience temporary abundance. With vast areas of soil irreparably or almost irreparably ruined through unwise exploitation, and with fuller knowledge of the causes of this destruction, soil conservation becomes as important an aim as the expansion of agricultural production. Soil deterioration may be rapid or gradual, but immediate gains in food supply are dearly bought if they lead to the ultimate impoverishment or ruin of the medium from which they are obtained. The first question to be asked by those responsible for the agricultural policy of any country is, therefore, whether the use made of the soil is such as to ensure that this soil shall continue to be capable of yielding reasonable returns for the labour of the cultivator. The existence of the Soil Conservation Service in the United States is evidence that the maintenance of the soil is a major objective in the nation's future agriculture. Other countries also have become unceasingly aware of the importance of conserving their soil.

In western Europe and particularly in Britain, the equable character of the climate and the prevalence of rational and conservative systems of land utilization render the fear of soil destruction fairly remote. Here the danger is rather that the potentialities of soil and climate are not realized to the full. Anyone familiar with practical farming knows that the difference in output between land poorly farmed and the same land well farmed is very great indeed. In the experience of the writer the adoption of "good husbandry" on Welsh farms in place of "hand to mouth" farming has resulted in an approximate doubling of production as measured by stock-carrying capacity. Without any revolutionary changes in farming practice, a general grad-
ing up of the standard of husbandry, including the improve-
ment of the vast areas of marginal grazing, might produce a
striking increase in food production from British farm lands.

The destruction of soils under cultivation, though
operative only in certain regions, is, for the future, a
menace to the world supply of food and other commodities
obtained from the soil. Much of the damage done in past
generations is well-nigh irreparable, and the prevention of
further losses is one of the main tasks confronting the
pedologist. Hitherto, the practical aims of soil investiga-
tion have been chiefly concerned with the increase of crop
growth. In future, the problems of soil conservation must
claim more attention, particularly in view of the potential
developments of tropical agriculture. These problems
demand the most ample elucidation of the constitution and
properties of the soil. The successes that have been
obtained encourage the hope that the fullest utilization of
the world's soil resources may not be incompatible with the
conservation of those resources by systems of permanent
agriculture. There shall be an handful of corn in the earth
upon the tops of the mountains (Ps. lxxii, 16).

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