SOIL GENESIS AND CLASSIFICATION

SIXTH EDITION

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S. W. BUOL
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R. C. GRAHAM
P. A. McDANIEL

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Preface to the First Edition

We have intended in this work to summarize the body of knowledge called pedology, to direct readers to sources of additional information in the literature, and to encourage students to learn directly from the soil in its natural setting. This book is one in a succession of periodic reviews of soil morphology, genesis, and classification that may serve as stepping stones across the seeming morass of terminology and information.

These are exciting times in pedology. The “information explosion” in the several fields of soil science has enabled us to better understand soils and define them more quantitatively. The adoption and use of the Comprehensive Soil Classification System has provided new concepts and nomenclature. And the uses of soil survey, the end product of our classification, have greatly increased through their interpretation for application to land use and productivity studies, especially for nonfarm land use problems.

Yet the student, formally enrolled in college or a self-taught learner, has not had access to any summary in the form of an up-to-date reference or text. We have designed this book with the hope and desire that it will be equally useful to graduates and advanced undergraduates, professional pedologists and geographers, ecologists, and all others interested in or involved with the land. Portions of this book we hope and expect will be of use to planners, to highway engineers, and to sanitarians involved with disposal problems. An understanding of soil genesis and classification is a prerequisite to sound land use planning and land management. Soil science can help people learn to live adequately and significantly in a varied ecosystem and derive necessities from it without damaging it.

Limited use has been made of sketches to illustrate soil profiles. It is our intent that these be supplemented by color slides to illustrate classroom lectures. The Marbut memorial collection of 2-by-2-inch color slides assembled by the Soil Science Society of America is excellent for this purpose.

Our effort is in appreciation of the direction and motivation of which we have been the fortunate recipients in the past and an attempt to share what we have learned. We are grateful to the many persons who helped us in preparing this manuscript. The responsibility for any oversights is ours.

Madison, Wis., 1973

S. W. Buol
F. D. Hole
R. J. McCracken

Preface to the Second Edition

The second edition has entailed considerable revision that includes updating those portions that deal with Soil Taxonomy, which had not been formally published at the time of preparation of the first edition. We have also updated the soil classification systems of other countries and of FAO according to revisions and additions that have been made since the early 1970s. Chapters 25 and 26 have been completely rewritten, consistent with what we believe to be desirable trends in considering soils in three dimensions (as soilscapes) and in increasing interpretation of soil maps and scientific comprehensive soil classification systems for many specific technical applications. The recent publication of *Minerals in Soil Environments* by the Soil Science Society of America (J. B. Dixon and S. B. Weed, eds.) has greatly added accessibility to soil mineralogical literature. The experiences of the authors since 1973, as well as comments from students and other readers, have led to the removal of some ambiguous and erroneous statements and to the addition of some new material in most of the chapters.

Madison, Wis., 1978

S. W. Buol
F. D. Hole
R. J. McCracken

Preface to the Third Edition

The third edition reflects an update of literature and experience during the past seven years. More than that, it attempts to reflect the "coming of age" of *Soil Taxonomy* as a major unifying force in promoting worldwide communication and understanding of soil morphology, genesis, and classification. Since our preparation of the second edition, considerable effort and progress have occurred in bringing global experience to *Soil Taxonomy* through the formation of international soil classification committees and through the publication of many papers and symposia. The efforts of these committees resulted in major changes with minimum violence to the system. Because of the official recommendations of the international committees, we feel that *Soil Taxonomy* has now accomplished the most difficult of its objectives, namely, that "the taxonomy should be capable of modification to fit new knowledge with a minimum of disturbance" (Soil Survey Staff 1975, p. 8). Great progress has also been made toward another of its goals that "the taxonomy should provide for all soils that are known, wherever they may be" (Soil Survey Staff 1975, p. 8). Therefore, in this edition we have attempted to adhere more closely to terminology and provisions of *Soil Taxonomy*, updating the terminology of older literature where necessary and stressing more recent research results. We have included discussions of approved amendments to *Soil Taxonomy*, such as the Kandi great groups, as well as pending proposals for major changes in the classification of Oxisols and the proposed Andisol order.

We are mindful, and stress to our readers, that classification should neither bias nor limit investigations concerning soils. Students should be encouraged to challenge the "wisdom of the past" that is the basis of any classification. In soil genesis, it is often said that the experiments have been conducted. Thus, the soil scientist is left to observe the result and interpret the processes by which the pedologic transformations have taken place. As such, the systematic and orderly observation of soil in the field, with the assistance of analytical techniques on samples brought back to the laboratory and greenhouse, remains the basis for future advances in soil science.

In this third edition, we have updated our review of soil classification systems used in other countries so readers may fully appreciate how soils are viewed and classified around the world.

We also have added new material reflecting the advances in soil genesis (that we have brought into chapters on flux factors and site factors) and in soil cover (including soils and landscapes and spatial variability). And the new developments and approaches in soil survey interpretation are reflected in a revised and expanded chapter on that subject.

1988

S. W. Buol
F. D. Hole
R. J. McCracken

Preface to the Fourth Edition

This fourth edition reflects advances in concepts and new information based on research and field experience. It also describes the many changes in soil classification reflected in *Soil Taxonomy* in the six years since the third edition was prepared and distributed. This edition reflects several changes in terminology, advances in technology, and our increased understanding of soil formation processes.

We draw attention to the reorientation and expansion of the range and scope of coverage in soil genesis and classification. This includes additional emphasis on the biogeochemistry of soil systems and the roles of soils in ecosystems.

This edition includes for the first time a definition and description of soils in the coldest regions of the planet that are actively being considered as a new order, Gelisols, in *Soil Taxonomy*. Also, in this edition you will find more information about the genesis, properties, and classification of soils frequently present in tropical and subtropical regions and soils with andic soil properties, the Andisols.

Also included are discussions and explanations of the use and applications of new technology via soil databases, soil geographic information systems, ground penetrating radar, increased use of aerial photography for soil studies, and new approaches to modeling soil systems. We also discuss the growing problem of trying to characterize and adjust for spatial variability within soil mapping units. Also included in this edition are discussions of concerns about soil quality, measures for remediation of soil problems such as waste disposal, and soil loss by erosion. Certain chapters of the previous editions have been combined and blended to produce a more comprehensive and cohesive picture and description of soil systems.

Another plus is the addition of a fourth author who is actively teaching and conducting research in arid and semiarid areas as well as internationally.

1997

S. W. Buol
F. D. Hole
R. J. McCracken
R. J. Southard

Preface to the Fifth Edition

This fifth edition presents the current understanding of soil as a thin layer of the earth's surface that forms an interface between the inorganic minerals of the earth's crust and the organic components of biologic entities living on the land and in the surrounding atmosphere.

Soil genesis, or the more encompassing concept of pedology, is an evolving discipline devoted to understanding how soils differ in form and function. The widely held view that a soil is a formed and therefore a nearly stable entity has evolved into the view that soil is a dynamic natural entity that interfaces with ecosystems and human endeavors. Chemical, biological, and physical reactions are constantly taking place within soil. Some of these reactions produce changes so slowly that human time scales are inadequate for measuring and comprehending soil dynamics. Other changes are rapid, episodic events that are viewed by humans as catastrophic and that challenge soil scientists to devise techniques to quantify the magnitude and probability of their occurrence.

Human efforts to understand and communicate knowledge about soils are reflected in attempts to classify soils. This edition reviews a spectrum of soil classification systems and presents the culmination of more than two decades of testing and revisions in the most detailed and comprehensive system for soil classification as reflected in the 1999 publication of the second edition of *Soil Taxonomy*. To better reflect the state of knowledge, new nomenclature has been added and improvements have been made in the systematic structure of soil classification categories. New analytical techniques have been utilized to more quantitatively identify soil properties and define class limits.

While we accept that soil changes, as external energy from the sun and rain is dissipated on its mineral components and as new organic components are added from the biological organisms that invade, we have to accept that humans also change. Just as in soil, where new components are formed to replace weathered components, new authors have replaced some of the stalwart authors of earlier editions.

2003

S. W. Buol
R. J. Southard
R. C. Graham
P. A. McDaniel

Preface to the Sixth Edition

This sixth edition presents current concepts of how soils are formed and sustained as an entity on the surface of the land masses on planet earth. Soil is an entity composed of mineral material inherited from geologic rock as altered by the actions of water and additions of organic materials injected by the plants and animals that find root within and tread on its surface. In response to the multitude of geologic materials, climatic conditions, and biologic ecosystems that presently exist and existed in the past, soils have acquired a multitude of properties that defy attempts to characterize soil as a singular entity. Thus, we speak of soils of the earth.

Human efforts to better understand soil properties and behavior and to more efficiently utilize soils for human endeavors have resulted in numerous attempts to classify the various kinds of soil. This is a never-ending endeavor that proceeds as new chemical, physical and biological methods are developed to analyze soil material. There is also the realization that soils have daily and seasonal dynamic features of temperature and moisture that affect human usage. All of these have to be considered and addressed in the search for more comprehensive understanding of the role soils play in ecosystem function, and how humans can best utilize and conserve the various kinds of soil.

This edition builds on the material contained in earlier editions and incorporates more detailed data regarding specific examples of how soils acquire various characteristics. We include a section of color plates that help readers better visualize the various kinds of soil. Also included is the recent recognition that soils periodically covered by shallow water are an important component of the spatial association between land and sea. We have expanded on methods of communicating and analyzing spatial information that have been greatly enhanced by modern electronic technologies. These technologies enable the rapid dissemination of information among soil scientists and users of soil information, and drive the development of new methods for modeling processes of soil formation and resulting patterns of soils on the landscape.

Our knowledge of the soils in the world is ever increasing as we utilize new analytical methods to investigate soil chemical, physical, and biological properties from the nano to the landscape scale and modern methods of transportation to gain access to previously little-studied areas of the world. This has greatly expanded not only our understanding of how soils are formed but also the techniques different human cultures utilize to manage the various kinds of soil from which they obtain human nourishment.

It is through the amalgamation of a multitude of individual observations as reported in scientific publications that this edition has been assembled.

S. W. Buol
R. J. Southard
R. C. Graham
P. A. McDaniel

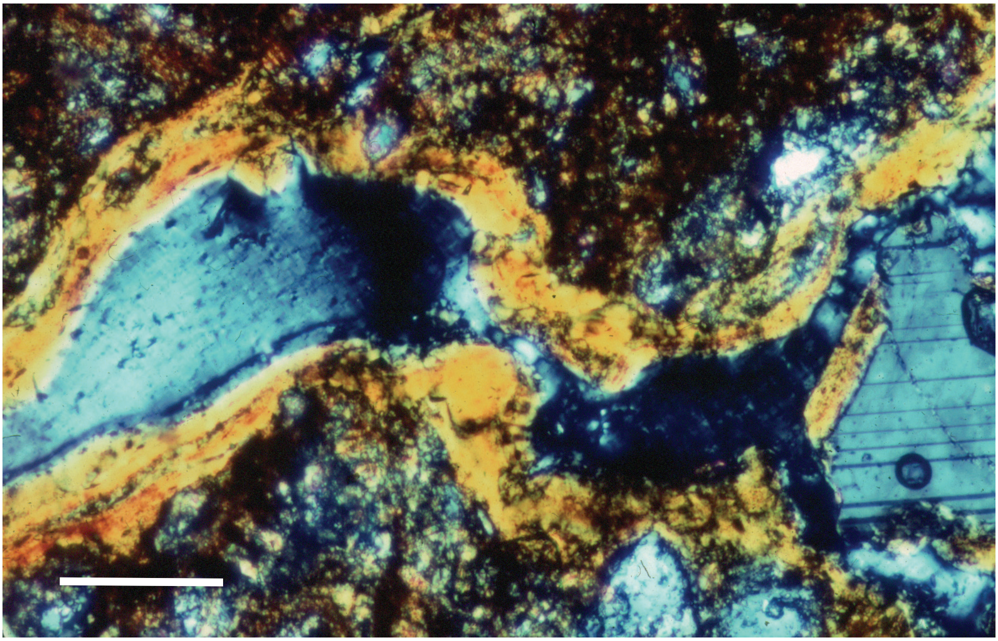

Plate 2.1. An argillan lining a pore in the argillic horizon of a California Palexeralf. Photomicrograph with crossed polarizers. The white bar represents about 250 μm.

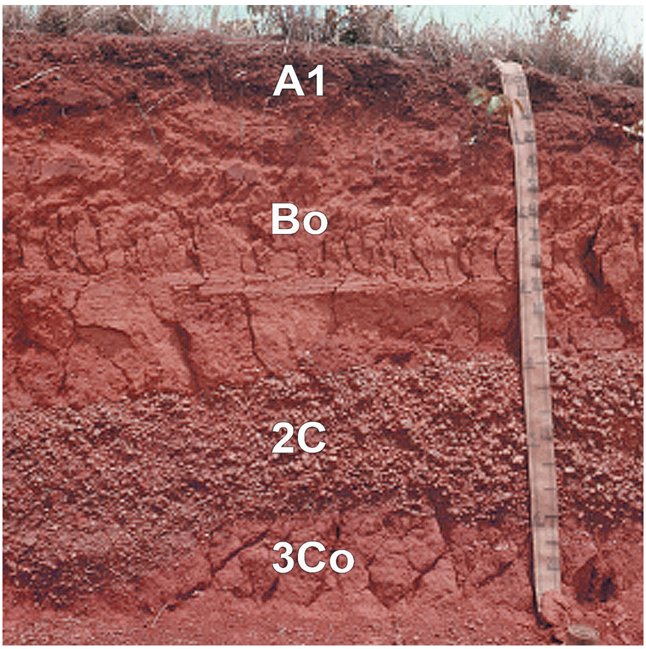

Plate 3.1. Profile of oxidized sediments in Federal District of Brazil. Note layer of ironstone and quartz gravel. Also note blocky structure appearance is due to road bank exposure. True structure of all horizons is granular.

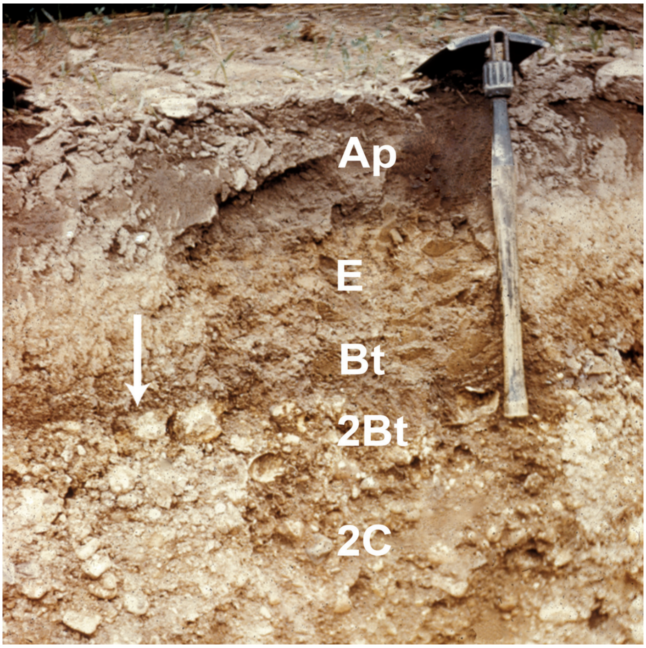

Plate 3.2. Soil profile developed in loess over till in Southern Wisconsin, USA. (Arrow points to contact between loess and till.)

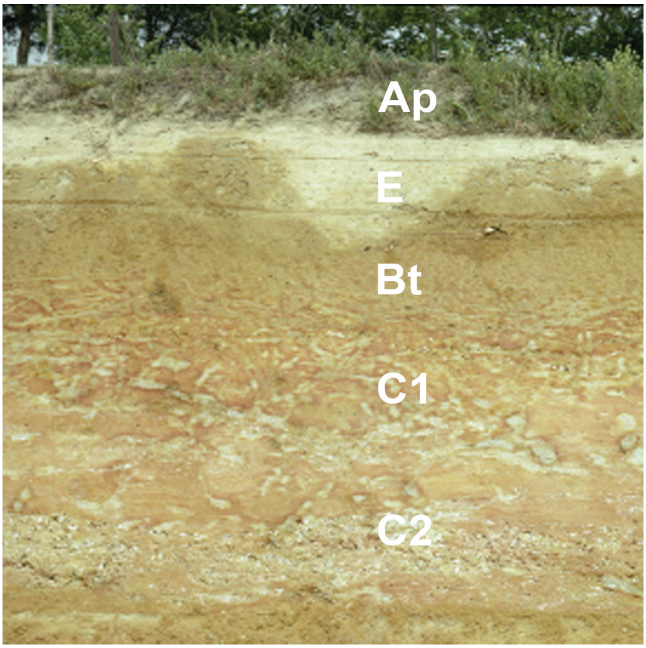

Plate 3.3. Road bank exposure of profile variations caused by uprooting of a tree. Site in Eastern North Carolina where an uprooted tree moved kandic horizon material, right to left with deepened E horizon at former tree site. All wood has decomposed.

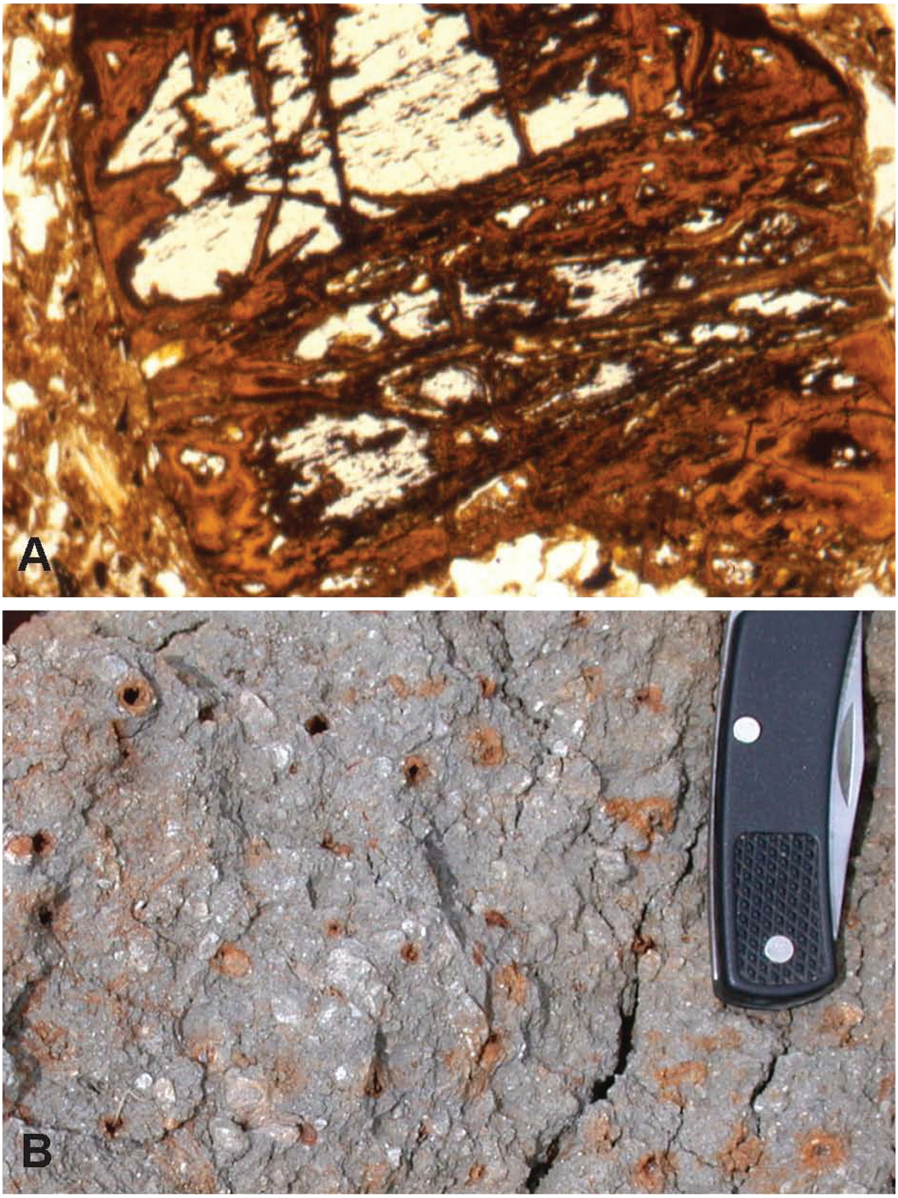

Plate 4.1. (A) Photomicrograph of an Fe-rich garnet (almandine) weathered to produce iron oxides (goethite and hematite). The yellow-brown iron oxides are concentrated in etch pit networks throughout the remnant grain, which appears white in the transmitted plane polarized light. (B) Photograph showing orange mottles of lepidocrocite produced by oxidation around root channels. The reduced soil matrix exhibits the low chroma (gray) color of its component silicate minerals.

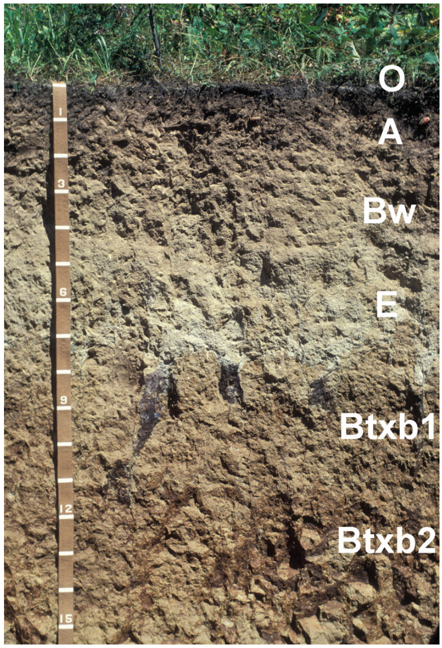

Plate 8.1. Vitrandic Fragixeralf (Santa series) from northern Idaho. Soil has formed in loess and reworked silty sediments and supports forest dominated by grand fir (*Abies grandis*). A well-developed fragipan is present a depth of ~80 cm. Scale is in decimeters.

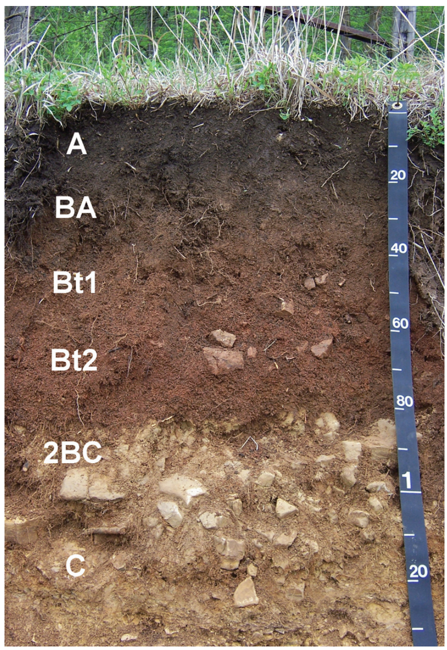

Plate 8.2. Mollic Hapludalf from Iowa County, Wisconsin. Soil has a mollic epipedon and argillic horizon; it has been leached free of $CaCO_3$. (Image courtesy of Dr. Randy Schaetzl, Michigan State University)

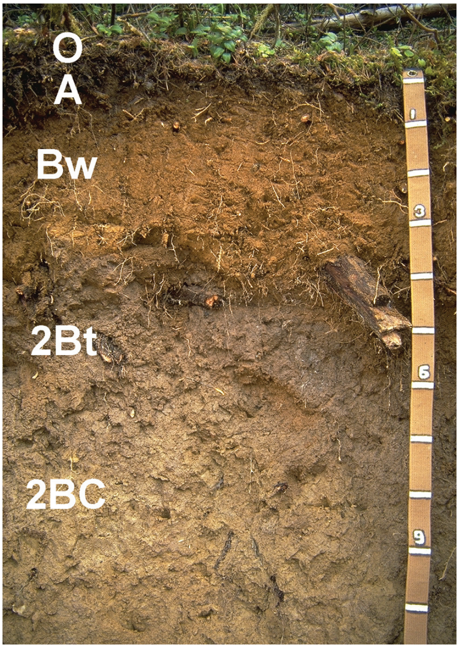

Plate 9.1. Alfic Udivitrand from northern Idaho. Soil has formed in 7,700-year-old ash from the cataclysmic eruption of Mount Mazama (now Crater Lake, Oregon). Volcanic ash mantle extends to a depth of ~40 cm. Scale divisions are decimeters.

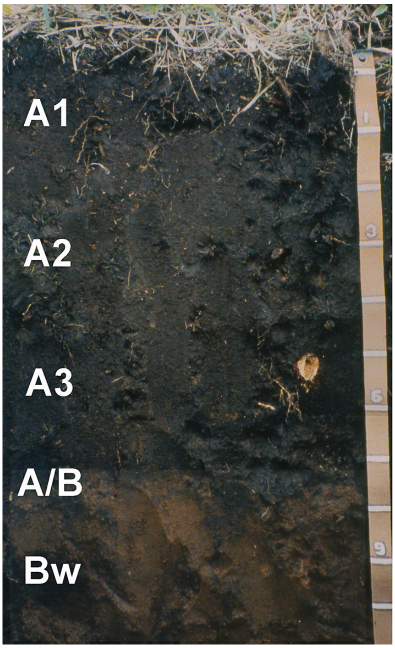

Plate 9.2. Melanudand from Costa Rica. Soil has a melanic epipedon and is an example of a nonallophanic Andisol in which organo-metal compounds dominate the colloidal fraction. Scale divisions are decimeters.

Plate 10.1. Photograph of a fine, smectitic, thermic Natric Petroaragid near Laguna Chapala, Baja California. Note the vesicular horizon (0–7 cm) under a desert pavement, the natric horizon (7–70 cm), and calcic horizon (15–70 cm). Laminar petrocalcic material caps the boulders in the subsoil.

Plate 10.2. Photograph of a coarse-loamy, mixed, superactive, thermic Typic Petroargid near Las Cruces, New Mexico. Note the platy structure of the petrocalcic horizon beginning at the 150-cm depth.

Plate 11.1. Photograph of a sandy-skeletal, mixed, mesic Typic Xerorthent formed in a 75-year-old debris flow deposit in the San Bernardino Mountains, California (Turk et al., 2008). Note the O horizon formed by the accumulation of organic matter on the surface, a weak A horizon in the upper 10 cm, and lack of other distinct horizons. (Photo by J.K. Turk)

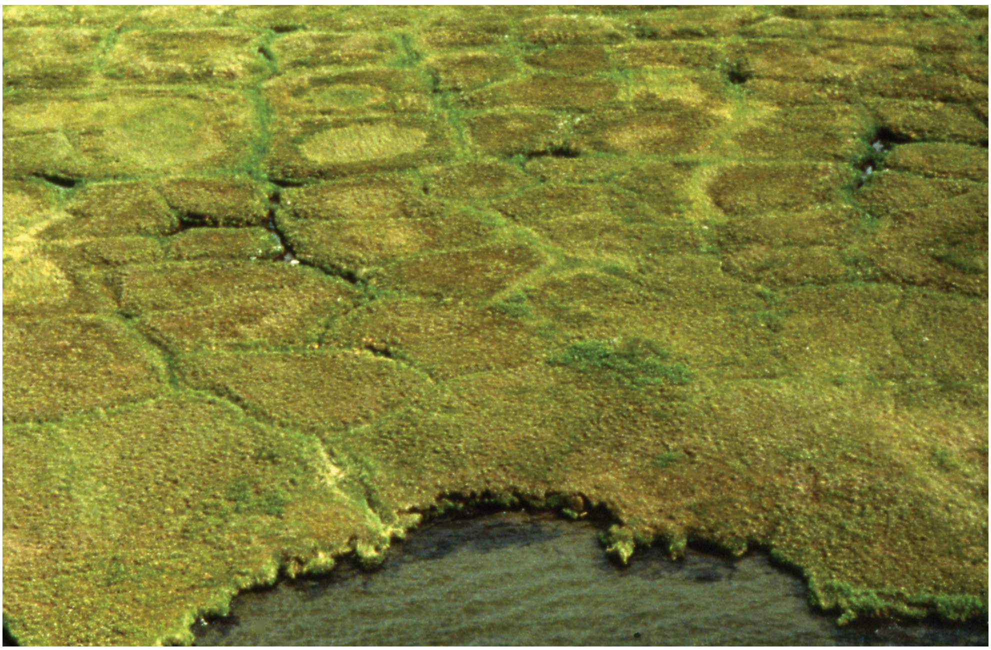

Plate 12.1. Gelisol landscape near Prudhoe Bay, Alaska showing patterned ground. Polygons are approximately 10 m across. (Image courtesy of Dr. Maynard Fosberg, University of Idaho)

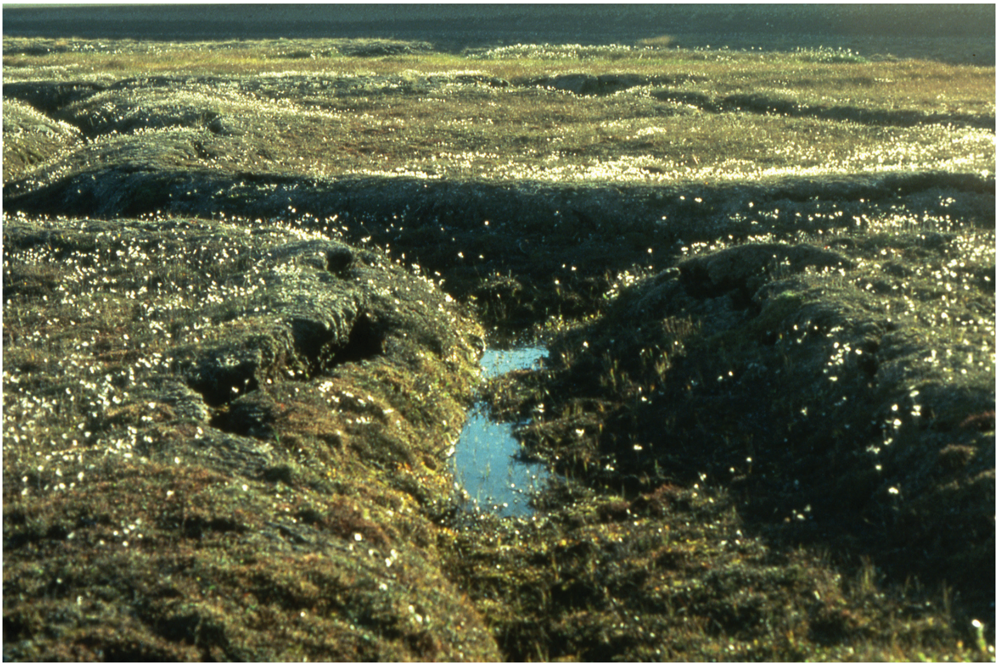

Plate 12.2. Close-up view of polygons shown in Plate 12.1 and are delineated by troughs up to 1 m deep. (Image courtesy of Dr. Maynard Fosberg, University of Idaho)

Plate 13.1. An alpine Cryohemist in Switzerland, formed mostly from sedges, with gray strata of mineral soil material. The black bar represents 50 cm. The water table is at a depth of about 80 cm.

Plate 14.1. Photograph of a Vitrandic Haploxerept near Craters of the Moon National Monument and Preserve, Idaho. Parent materials are mixed ash/cinders and loess. The mixed parent material, young age (~2,000 years), and relatively cool, dry climate have impeded development of andic soil properties so the soil is an Inceptisol rather than an Andisol. Note the 5-cm-diameter krotovinas in the lower part of the profile filled with yellowish brown material from the upper horizons. This is evidence of mixing by burrowing rodents. Scale is marked in 10-cm increments.

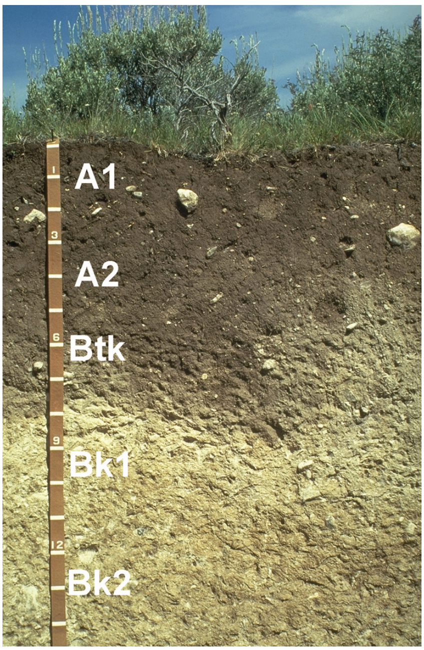

Plate 15.1. Pachic Argicryoll from Lemhi County, Idaho. Soil has formed in glacial drift and has a thick (pachic) mollic epipedon. Mean annual precipitation is ~430 mm; native grasses and sagebrush are the dominant vegetation. The upper right-hand side of the profile has been extensively mixed by badgers.

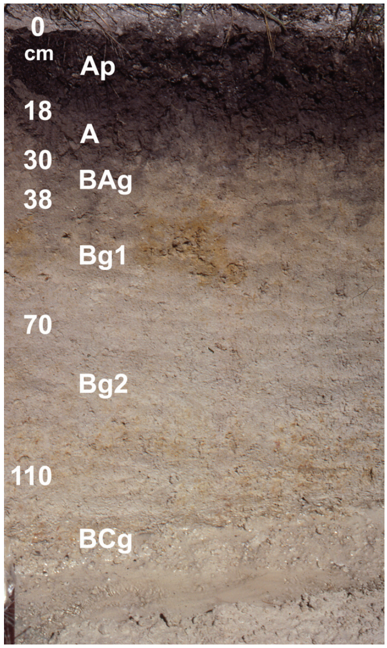

Plate 16.1. Photo of a Fine, kaolinitic isohyperthermic Aeric Haplaquox profile in the Federal District of Brazil. (Characterization data from this site is published on page 658 of *Soil Taxonomy* (Soil Survey Staff, 1999).

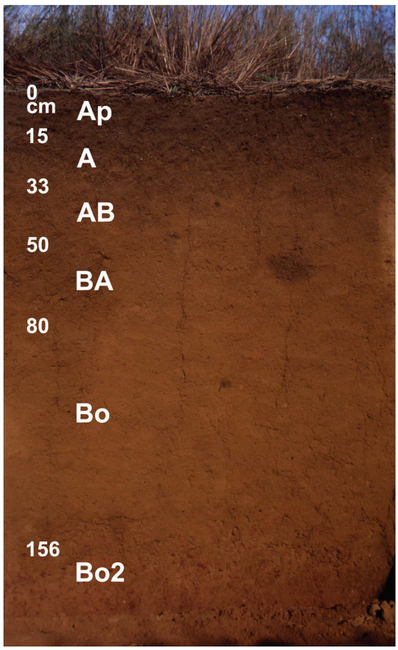

Plate 16.2. Photo of a very-fine, gibbsitic, isohyperthermic Typic Acrustox profile located in the state of Goiás, Brazil.

Plate 17.1. A Haplorthod in the boreal forest of southwest Sweden formed in loess over gneissic till. The pedon is about 1 m deep.

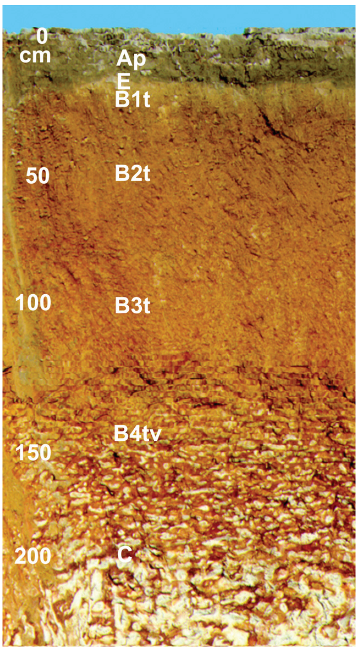

Plate 18.1 Photo of a Dothan (fine-loamy, kaolinitic, thermic Plinthic Kandiudults) profile formed in coastal plain sediments in Johnston county, North Carolina, USA.

Plate 19.1. A Hapludert in Texas. The dark-colored microbasin is on the left; the lighter-colored diapir on the right creates a microknoll. The scale on the left is in decimeters and feet. (From Eswaran et al. 1999)

SOIL GENESIS AND CLASSIFICATION

SIXTH EDITION

Introduction

1

This book discusses the composition of soils as natural bodies resulting from biogeochemical processes on the land surfaces of earth. It also examines human attempts to better understand the interaction of soils with biological components of the ecosystem, including humans, via the study and classification of soils. The main themes of soil genesis and classification follow:

1. Identification and description of soil profiles and pedons (soil morphology);
2. Characterization of chemical, mineralogical, and physical soil properties aided by laboratory and field investigations of soil properties;
3. Categorization and classification of soils according to similarity of properties and function;
4. Mapping the spatial distribution of soils as they exist on the earth's surface;
5. Analyses of the relationships between soil properties and the many potential uses of soils.

All of these activities comprise a recognized area of specialization within the discipline of soil science. The Soil Science Society of America changed the name of its Soil Genesis, Morphology, and Classification division to Pedology about a decade ago (Simonson 1999). "Pedology" (from Gr. *Pedos*, "ground," and *logos*, "science"; original formed as Russian, *pedologiya*) is a collective term used to refer to the combination of the two phases of soil science: (1) soil genesis and classification and (2) more inclusively, also soil morphology, survey or mapping, and interpretations. Pedology is practiced in many other countries, especially European and Asian countries and Australia (Editorial Staff 1940; Gibbs 1955; Leeper 1953, 1955; Northcote 1954). The International Union of Soil Sciences includes these aspects of soil science primarily in their commission of Soil in Space and Time.

Subdivisions of Pedology

Following are descriptions of each of the distinctive phases of activity encompassed by pedology.

Soil genesis is a century-old science that has dealt with soil in three conceptual phases: (1) as a geologic entity, (2) as a product of factors and processes of soil formation, and (3) as an open system capable of supporting the functions of soil in all

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

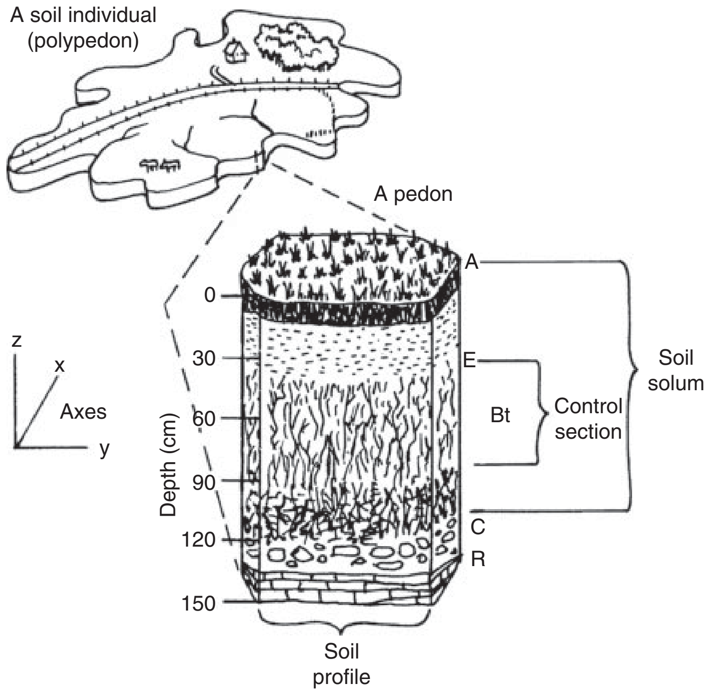

Figure 1.1. A soil individual is a natural unit in the landscape, characterized by position, size, slope, profile, and other features.

ecosystems. Soil genesis includes concepts of biogeochemistry. It conceptualizes the factors and processes responsible for the chemical, physical, and mineralogical properties of all soils and the spatial distribution of various kinds of soil on the landscape.

Soil classification is the categorization of soils into groups at varying levels of generalization according to their physical, mineralogical, and chemical properties. The objectives of soil classification include organization of knowledge, ease in remembering properties, clearer understanding of relationships, and ease of technology transfer and communication.

Many classification systems are used to classify soils and soil materials. Some are designed to relate soil properties to specific uses and are referred to as "technical classification systems." Others, termed "natural classification systems," are structured to categorize all soil properties. The primary classification system used in this book is *Soil Taxonomy*, a natural classification system. When the term *Soil Taxonomy* is capitalized and italicized in this book, it refers to the system of classification developed by the U.S. Department of Agriculture (USDA) Soil Survey Staff with the support of and contributions from land-grant universities, other federal and state resource agencies participating in the National Cooperative Soil Survey, and colleagues from other countries. It was formally released and published by the USDA in 1975, and a second edition was published in 1999 (Soil Survey Staff 1975, 1999). The system is also updated periodically through the release of revised *Keys to Soil Taxonomy*. The most recent edition of the *Keys* was published as the 11th edition (Soil Survey Staff 2010). *Soil Taxonomy* is now widely used throughout the world. *Soil Taxonomy* uses quantitative morphological criteria to define kinds of soil and concepts of soil genesis to guide the selection and orderly application of these criteria.

All classification systems are a mirror or state-of-the-art indicator of the available knowledge about the objects classified, assembled in a systematic fashion to facilitate communication with other scientists and, most importantly, with the students who will become the future scientists. The fact that classifications systems change over time is an inevitable result of research in soil science and in other sciences that relate to uses of the soil.

Soil morphology encompasses the color, physical structure, chemical and mineralogical properties of soil material; the spatial association of materials in soil horizons (Figure 1.1); and the temperature and moisture dynamics of soil in situ. The thickness, vertical relationship, number, and three-dimensional ranges and variations of horizons found in the smallest recognizable volume of soil that is classified (called the "pedon," the word rhymes with "head-on") are described and recorded by a standard nomenclature outlined in Chapter 2.

Perspective on the Role of Soil Genesis and Classification

It is useful, important, and interesting to consider how the study of soil genesis and classification interacts with other fields of soil science and other scientific disciplines. This is especially true for technical soil classifications derived from scientifically based natural classifications systems like *Soil Taxonomy* (Buol and Denton 1984). Soil properties are primary reagents in field experimentation. Documentation of soil properties at research sites is essential for the successful transfer of research results to other locations.

Soil genesis and classification studies have made contributions to research design and data acquisition in other fields of soil science, including biogeochemical redistribution of nutrients in ecological systems, ecology of soil microbes and mycorrhizae, and the availability and distribution of plant essential nutrients such as phosphorus and nitrogen in different types of soils (Runge and McCracken 1984). Soil maps furnish basic inputs to soil conservation planning in the United States and provide information used in equations for predicting soil loss and water pollution potential under various management practices on different soils.

Characterization of soil properties is fundamental to all soil studies. Complete soil characterization for classification purposes requires that all horizons of the soil be analyzed. Many laboratory and greenhouse studies require only characterization of soil material from a few horizons, but practical application of the results obtained from such studies requires verification with field studies. Soil characterization methods draws heavily on methods of soil chemistry, physics, mineralogy, microbiology, and biochemistry. Conversely, methods and results initially obtained in soil characterization and soil genesis studies have been useful in perfecting methods of soil analysis by providing materials representative of all kinds of soil for analysis.

Soil genesis embraces the concept of soil as "a natural entity to be studied as a thing complete in itself" (Cline 1961). This concept has survived the fragmentation of soil science into the subdisciplines of soil chemistry, soil physics, soil microbiology, soil fertility, and soil management by drawing upon and integrating the concepts, theories, and facts of these subfields of soil science into a holistic, integrated, multidisciplinary view of soil as a natural entity. Soil genesis and classification, or pedology, may also be likened to a system of bridges connecting the disciplinary islands of geology, biology, chemistry, physics, geography, climatology, agricultural sciences, economics, anthropology, and archeology. The interdisciplinary nature of this field of soil science gives it added importance in the training of scientists (Abelson 1964). Soil genesis and classification and its allied activities therefore have many interactions with and contributions to fields of science other than the science of soils.

Soil genesis and soil classification have some roots in geology, for they grew out of the study and mapping of rocks. The close ties between geology and soil science stem from the fact that most soils are derived from geologic materials such as granite, limestone, glacial drift, loess, and alluvium. Several of the early pioneer soil scientists were geologists by training. Because differences among soils are due in part to the different landforms they occupy, and because age of the soil is related to the stability of the surface on which they have formed, close ties between soil specialists in genesis and classification and geologists specializing in geomorphology continue to be strong and mutually beneficial.

Soil genesis and classification is seldom concerned with entire geological deposits, but rather deal with the upper portion of the deposit that has been influenced by plant and animal activity and by the intrusion of water and energy from the land surface. Therefore, soil genesis and classification, which deals with the dynamic, biologically active soil system, must also be concerned with biology, especially the subsciences of ecology, microbiology, plant physiology, and botany. Hans Jenny (1980) regarded soil and vegetation as coupled systems and thus an ecosystem. A knowledge and awareness of plant-soil interactions, meteorology, and hydrology are essential for soil scientists interested in soil genesis and classification.

Soil underpins human food production and is a very significant component in our total stock of natural resources. Production economists call on soil scientists for data and estimates of the productivity of various soils under defined systems of management. Natural resource economists are concerned about the amount and distribution of useful productive land. Planners at local, state, regional, and national levels use soil surveys and soil interpretations for land-use planning, environmental protection, selection of building and highway sites, tax assessment, and land evaluation (Simonson 1974; Bauer 1973; Jarvis and Mackney 1979). This places additional responsibility on soil classifiers and interpretation specialists to provide sound, scientifically based evaluations of flooding, structural instability, and other economic potentials of individual kinds of soil. Soil scientists must consider economic and resource conservation factors in the preparation of technical and natural classification systems. Archeologists and anthropologists utilize soil information and data to date construction and destruction of human settlements and explain changes in agricultural practices (Olson 1981; Harrison and Turner 1978).

It is clear from the preceding examples that soil genesis and classification interact with a large number of disciplines and interests. Not only does this signify the important and varied uses for soil information, but it also challenges those working in this area to be aware of all potential uses as they seek to better understand the dynamic role of soils in the environment.

Developmental Stages of a Discipline

Three stages in the development of a discipline enumerated by Whitehead (1925) may be applied to soil genesis and classification and their related components:

Stage 1: Location in space and time. The basic operation of mapping soils to record their positions in space is a prerequisite to genesis and classification and their related components.

Stage 2: Classification. Whitehead (1925) calls this a "halfway house between the immediate concreteness of the individual things and the complete abstraction of mathematical notions." A great variety of genetic and descriptive soil classifications have been developed. Terminology used ranges from symbols, to synthetic terms based on classical languages, to a hodgepodge of folk terms. All classification systems rely on the knowledge and understanding available and are subject to change as that knowledge and understanding evolve.

Stage 3: Mathematical abstractions. More highly developed abstractions are possible by mathematical means. Mathematical models also help us to predict. Relationships between soils and other natural phenomena are conceived from observations. Statistical expressions are possible when sufficient quantitative data become available (Bidwell and Hole 1964; Hole and Hironaka 1960; Arkley 1976; Jenny 1941, 1961b, 1980). Computer-based models that project the effects soil properties have on numerous entities utilize both soil characterization data and spatial data recorded on soil maps (Bouma 1994). Attempts to utilize existing data to model soil genesis and classification, and to aid soil mapping have become an increasingly important part of pedologic research (Bouma and van Lanen 1987; Vereecken et al. 1992; Tietje and Tapkenhinrichs 1993; Rasmussen et al. 2005; Hartemink et al. 2008).

Developmental stages in soil classification are not chronological. Although soil mapping has identified many of the soils in the world, many areas remain poorly explored. Also, the methods for identifying soil properties are improved as new analytical technologies are developed. Classification of an object such as soil cannot be considered complete until everything is known about that object. Classification can only reflect what is known, and in soil science this knowledge base increases as new lands are explored and new analytical techniques are employed to describe and analyze soils. Mathematical abstractions are built on a foundation of what is known. Improvements can be made as more extensive and intensive databases become available.

The components of soil science to which we have been referring are subjects in this book. We think it desirable to have a brief summary of the historical development and a synopsis of the principles, concepts, and theories plus notes on the methods of each before proceeding to the specifics.

Historical Developments in Soil Genesis

By reviewing history we gain perspective about modern soil genesis concepts, become aware that this field of science is dynamic, come to appreciate the resistance new ideas have encountered, and become aware of the newness of soil science as we know it today.

Aristotle (384–322 B.C.) and his successor Theophrastes (372–287 B.C.), considered the properties of soil in relation to plant nutrition. Roman writers who discussed differences among soils in relation to plant growth included Cato the Elder (234–149 B.C.), Varro (116–27 B.C.), Virgil (70–19 B.C.), Columella (about A.D. 45), and Pliny the Elder (A.D. 23–79). In 1840 Justus von Liebig published *Chemistry Applied to Agriculture and Physiology* in which he states that plants assimilate mineral nutrients from the soil and proposed the use of mineral fertilizers to fortify deficient soils (Liebig 1840).

In the middle of the nineteenth century several German scientists, including Ramann and Fallou, developed agrogeology, which viewed soil as weathered, somewhat leached surface mantle over rock. Fallou (1862) suggested that "pedology"—which to him signified theoretical geological soil science—be distinguished from "agrology," the practical agronomic soil science.

In Russia, Lomonosov (1711–1765) wrote and taught about soil as an evolutionary rather than a static body. In 1883, V. V. Dokuchaev (1846–1903) (Figure 1.2) published a report of a field study of Chernozem soils present under grasslands in which he applied the principles of morphology to soils, described major soil groups and their genesis, produced the first scientific classification of soils, developed methods of mapping soils, and laid the foundation for soil genesis and soil geography. In 1886, Dokuchaev proposed that "soil" be used as a scientific term to refer to "those horizons of rock that daily or nearly daily change their relationship under the joint influence of water, air, and various forms of organisms living and dead" (Vilenskii 1957). He later defined soil as an independent natural evolutionary body formed under the influence of five factors, of which he considered vegetation the most important. Russian soil scientists K. D. Glinka

Figure 1.2. V. V. Dokuchaev.

Figure 1.3. E. W. Hilgard.

(1867–1927) and S. S. Neustruyev (1874–1928) reemphasized the concept of soil as a superficial geological entity, a weathered crust that exhibits specific properties correlated with climatic zones. V. R. Williams (1863–1939), another Russian soil scientist, developed the concept of soil genesis as essentially a biologic process rather than a geologic one. He stated that soil formation takes place best in grasslands. P. E. Müller (1878) wrote a monograph on soil humus, describing the biological character of soil genesis in forests.

In the USA, E. W. Hilgard (1833–1916) (Figure 1.3), working as a geologist and soil scientist in Mississippi and California, wrote about the relationships between soils and climate (Hilgard 1892). He "saw the farmer's dirt as a richly embroidered mantle of earth, whose design and fabric were deserving of scientific zeal and quest" (Jenny 1961a). G. W. Coffey, a soil scientist with the U.S. Division of Soil Survey, published a soil classification system for the United States based on the principles of soil genesis expressed by Dokuchaev and Glinka (Coffey 1912). However, this was an idea whose time had not yet come in the United States, because the idea of soil as a superficial geologic material still dominated.

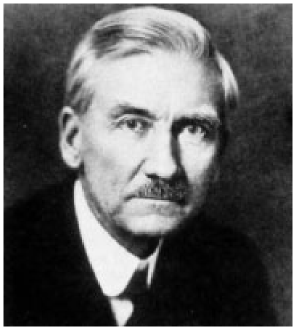

Figure 1.4. C. F. Marbut.

Figure 1.5. C. E. Kellogg.

C. F. Marbut (1863–1935) (Figure 1.4), while director of the U.S. Soil Survey Division, read Glinka's publications on soil genesis and classification and introduced their concepts into soil survey programs in the United States, as well as introduced his own emphases on the soil profile and the 'normal' soil (Krusekopf 1942).

Charles E. Kellogg (1902–1977) (Figure 1.5), Marbut's successor as director of the USDA Division of Soil Survey, continued the enhancement of the themes and principles of soil genesis as a necessary basis for soil classification and soil survey while introducing uniform techniques and nomenclature among soil scientists (Kellogg 1937; Soil Survey Staff 1951, 1960; Kellogg 1974).

Hans Jenny (1899–1992) (Figure 1.6) wrote a masterful treatise on the five factors governing the development of soils (Jenny 1941). He noted that quantitative elucidation of the processes of soil formation could not proceed without a larger body of data than was available. In a later work, Jenny (1980) put the five soil-forming factors

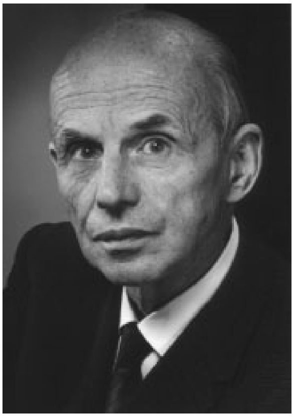

Figure 1.6. H. Jenny.

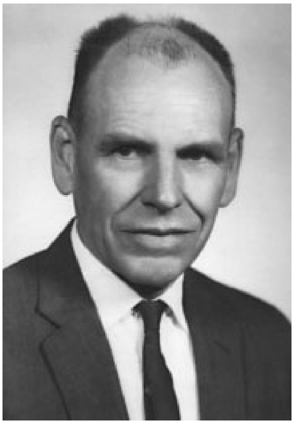

Figure 1.7. G. D. Smith.

"into a conceptual framework that permits solving the equation when landscape configurations are favorable."

In 1959, R. W. Simonson brought out the significant point, not well recognized previously, that many genetic processes may be simultaneously and/or sequentially active in any one soil (Simonson 1959). He pointed out that the horizons of a soil profile reflect the relative strength of these processes and the degree to which they offset each other.

In the 1960s and early 1970s, G. D. Smith (1907–1981) (Figure 1.7), as director of Soil Survey Investigations for the USDA Soil Conservation Service and chief architect of *Soil Taxonomy* (Soil Survey Staff 1975), with his colleagues, further advanced and refined soil genesis studies in support of soil classification and soil survey. Smith made the important point that concepts of soil genesis are very important for soil classification, but factors and processes of soil genesis cannot be used as the sole basis for soil classification because they can rarely be quantified or actually observed (Smith 1983). He stressed that soil classification was best based on properties that could be observed and measured within the present soil.

Concepts of Soil Genesis

Concept 1. Soil-forming processes, also referred to as pedogenic processes and biogeochemical processes, that are active in soils today have been operating over time and have varying degrees of expression over space (that is, at various locations). The geologic uniformitarian principle that states "the present is the key to the past" is also applicable in soils with respect to downward translocation, biocycling, and transformation of materials—back to the time of appearance of organisms on the land surface. We can improve our elucidation of soil formation of differing soil profiles by application of this principle.

Concept 2. Many soil-forming (and soil-destroying) processes proceed simultaneously in a soil, and the resulting soil properties reflect the balance of both present and past processes (Simonson 1959). Soil-forming processes are actually combinations of specific reactions that are characteristic of particular time spans and conditions.

Concept 3. Distinctive combinations of geologic materials and processes produce distinctive soils. Observable morphological features in a soil are produced by certain combinations of pedogenic processes over time. The degree to which a morphological feature is expressed is dependent on the intensity and duration of the process.

Concept 4. Five external factors provide the reactants and energy to drive the pedogenic processes within the soil. These factors are climate, organisms, relief, parent material, and time.

Concept 5. Present day soils may carry the imprint of a combination of pedogenic processes and geologic processes not presently active at that site. Therefore, knowledge of paleoecology, glacial geology, and paleoclimatology is important for the recognition and understanding of soil genesis.

Concept 6. A succession of different soils may have taken place at a particular spatial location as soil-forming factors changed. The soil surface is lowered by erosion and dissolution of soil material and elevated by depositions of soil materials and tectonics. In this respect, the volume of material examined as soil on the land surface actually changes in vertical space over time.

Concept 7. The time scale for soil formation is much shorter than the geologic time scale and much longer than the age span of most biological species. The vulnerable position of soil as the skin of our dynamic earth subjects it to destruction and burial by episodic geologic events. Few, if any, soils are older than Tertiary and most no older than the Pleistocene epoch. Succession of vegetative communities and human activities often alter soil properties over short spans of time.

Concept 8. Complexity of soil genesis is more common than simplicity. Some processes that influence soil composition are discontinuous or episodic and disrupt soil features formed by other processes that are rather continuous over time.

Concept 9. Soils are natural sites for clay mineral formation on land surfaces. Most primary minerals on earth crystallize from magma at high temperature and pressure, and in the absence of free oxygen. When exposed to lower temperatures and pressures, free oxygen, meteoric water, and organic compounds near the land surfaces, the primary minerals decompose by processes known as weathering. Some of the elements reassemble into new mineral structures of clay size. It is likely that the clay particles in the shale and other sedimentary rocks are products of mineral alteration in soil prior to erosion and deposition.

Concept 10. Understanding and knowledge of soil genesis is useful in soil classification and mapping, but scientific classification systems cannot be based entirely on inductive concepts of soil genesis. Processes operating within a soil can seldom be observed or measured and are subject to change over time, which renders quantification difficult if not impossible.

Concept 11. Knowledge of processes of soil formation is basic to understanding the impact of human use and management. Humans alter both the factors and processes of soil formation in their attempts to improve a soil's performance for specific purposes. Knowledge of pedogenic processes helps assure compatibility of human actions and ambient soil conditions.

A Soil as an Anatomical Specimen

Just as Louis Agassiz (1807–1873) taught his students to learn about fish by making accurate drawings of specimens of fish, so Dokuchaev, Hilgard, Marbut, Kellogg, and others have taught us to learn about soils by describing them carefully (Marbut 1935). The *Soil Survey Manual* (Kellogg 1937; Soil Survey Staff 1951, 1993) provides detailed guides to the scientific description of soil profiles (Cline 1961). Observations recorded in a soil description represent the state of the soil body and surrounding landscape at a particular hour on a particular day. An accurate description of the soil profile is documentation needed to relate data obtained from laboratory or greenhouse studies of soil samples to the soil in situ on the land.

A Soil as an Energy Transformer

Soils and the vegetative covers that live in the soil are "energy transformers," receiving radiant solar energy by day and reradiating to space during the night. When radiant energy is absorbed at the soil surface, it is converted to sensible heat and warms the soil. Each night heat is lost through radiation to the atmosphere. If there is a dense

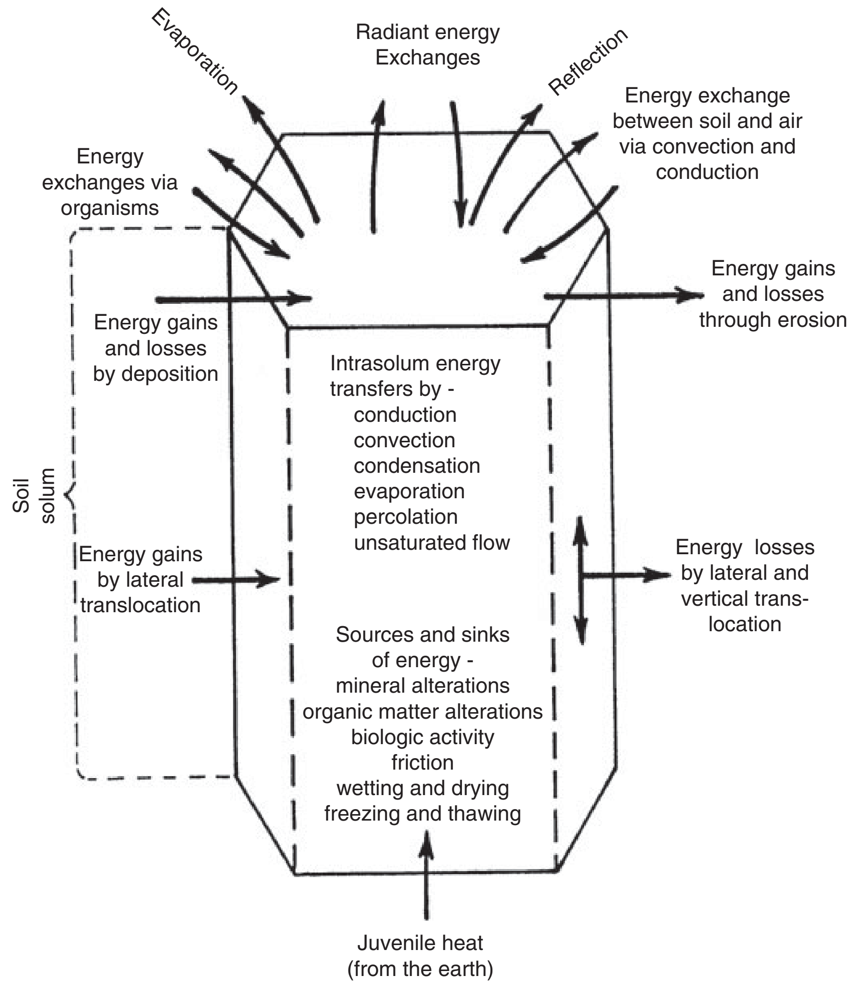

Figure 1.8. Schematic representation of energy sources and transformations in the solum.

cover of vegetation, the soil is shaded and experiences less heating and cooling. Small amounts of energy emanating from inside the earth (mostly as radioactive decay) escape as radiation from the soil. The energy transformations in the soil are accomplished through a variety of processes (Figure 1.8).

Each soil is fixed at a location and thus subject to the radiation dynamics at that location. Although the resulting soil temperature may be considered by some to be a climatic property, and not a soil property, soil temperature can be measured in the soil and is an important soil property that determines many of the interactions soils have with vegetation. Mean annual soil temperature and seasonal dynamics of soil temperature are used as criteria in *Soil Taxonomy*.

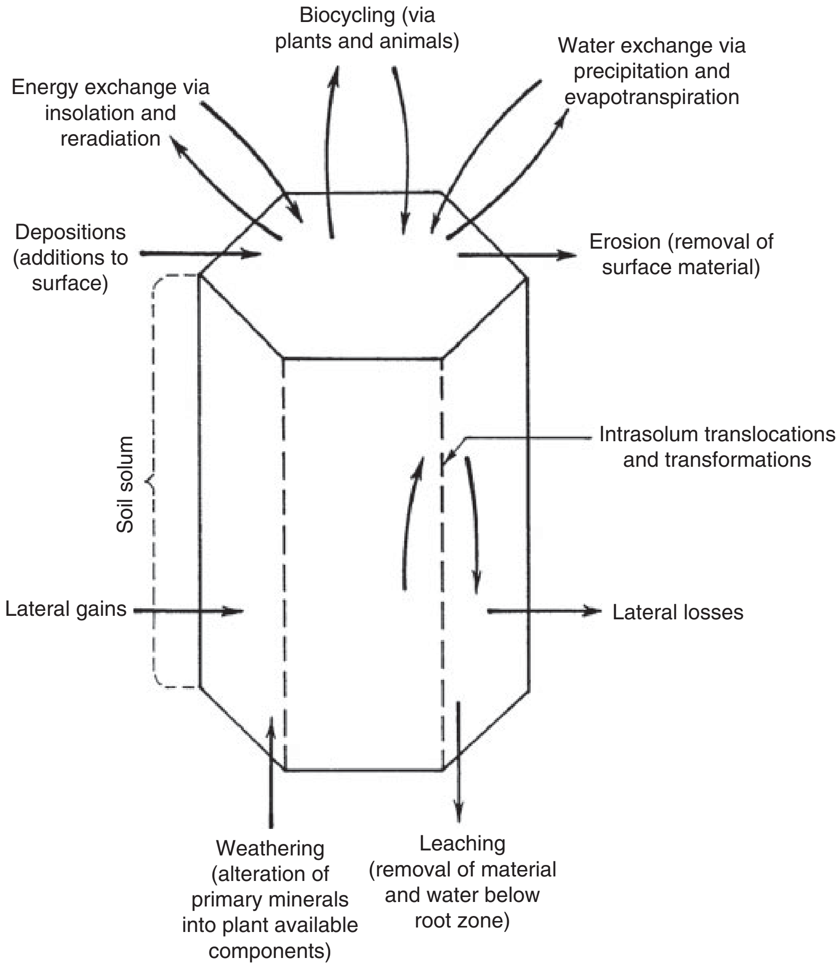

Figure 1.9. Schematic representation of the solum of a pedon as an open system.

A Soil as an Open System

A soil is an evolving entity maintained in the midst of a stream of geologic, biologic, and hydrologic activity (Figure 1.9). Individual soil bodies on the landscape and individual horizons within soil are the result of unequal distribution of materials within and between soils. Some soils and horizons within soils become enriched in certain substances; others become depleted. Water and energy flow through the soil surface, penetrate into the soil, and create an environment wherein primary minerals decompose and/or alter into secondary, mainly clay minerals (*weathering*). Plants and animals contribute to processes within the soil. The surface mineral horizon of a soil becomes enriched as carbon extracted from the air and chemical elements extracted from the entire rooting depth are deposited on the surface during leaf fall or death of the vegetation, whereas the lower part of the solum becomes relatively depleted in chemical elements extracted by plants (*biocycling*). There is transfer of material between soil bodies both at the surface (*erosion and deposition*) and by the subsurface flow of water (*lateral losses and gains*). Such transfers are more active on sloping land, wherein soils on the lower portion of a hill (footslopes and toeslopes) receive water and materials dissolved or suspended in that water from adjacent soils higher on the hillside (summits, shoulders, and backslopes). Materials may be removed from a soil in percolating water (*leaching*), but water from many rainfall events only percolates to a shallow depth, thereby translocating material dissolved and suspended in surface horizons to a subsoil horizon where it is deposited as plants transpire the water (*intrasolum translocations*) and new mineral and organic compounds may form (*intrasolum transformations*). Activity of earthworms, insects, macrofauna, the uprooting of trees, shrinking and swelling, and freezing and thawing all may act to reverse these downward translocations and physically move material upward within the soil.

Each soil has a budget of inputs and outputs that tends to establish a steady state at a given site. A steady state exists where the output is matched by an input of equal quantity and quality from a source not receiving the output. A steady state is often incorrectly considered equilibrium where material of equal quality and quantity moves to and from adjacent objects. Perhaps soils never reach a perfect steady state or equilibrium (Smeck et al. 1983), but in most soils the rate of change resulting from any imbalance is so slow it cannot be observed or measured. Where great imbalances exist, rapid changes in soil properties are observed. Perhaps the most noticeable changes are erosion losses on hillsides and depositional gains on floodplains.

The same factors and processes that influence soil formation influence terrestrial ecosystems. Processes of soil formation are components of ecological studies of potential global climatic and other environmental disturbances (Vitousek 1994).

Methods of Soil Genesis Study

Although there are many approaches to soil genesis study, depending on various circumstances, four general methods may be distinguished.

Independent Variable Method. This is a simple method insofar as it handles one item at a time (Jenny 1958). The intellectual assumption is made that all conditions are constant except one variable. For example, after installation of drainage to lower the water table in a soil rich in iron sulfide, the soil pH decreases to such a low level that little or no vegetation is able to grow. The obvious conclusion is that drainage has caused the soil to acidify. The danger in the independent variable method is that the observer may fail to verify that all conditions are the same and may be led to suppose that a relationship obtained in one place will hold elsewhere. The presence of iron sulfide is intrinsic to the result observed in the above example. When the soil is drained, air enters and oxidation reactions form sulfuric acid

(Fanning and Fanning 1989). It is obvious that drainage does not cause all soils to become strongly acidic or else we would have little agriculture as we know it today on the millions of hectares that have been drained.

Dependent-Variable Method. The soil complex is considered as a function of *n* variables, each of which can be written as a function of all the others, yielding *n* equations. This has the advantage of generalizing and dealing with a whole system. The results, on the other hand, may be quite unsatisfactory. For example, if the result of an analysis of a large soil region states that soils are extremely variable because conditions affecting the soils are extremely variable, no specific knowledge or understanding of soil formation is forthcoming.

Macroanalysis Method. By this method, the whole soil complex is divided into macro groups. This is actually a compromise between the oversimplicity of the independent variable method and the overgeneralization of the dependent-variable method. Examples of "macrogroupings" of soils are taxonomically defined soil groups, complexes, or associations of taxonomic groups. The risk in this method is that we tend to forget that taxonomic units are really complexes of individual soils, and we may consider them to be definite and stable entities.

Numerical Analysis Method. Numerical soil classification has been considered less objective and more fallible than traditional soil classification (Arkley 1976). Selection and weighting of soil properties interjects considerable bias into the analysis. Interest in numerical classification of soils was originally aroused by the success of the methods in terrestrial plant ecology, where a classification can be more objective than in pedology (Coventry and Williams 1983). In any case, numerical classifications of soils provide valuable insights. A necessity for numerical studies is access to data containing properties of soils (pedon databases) and geographic information systems that include data on soil-forming factors and processes (Anderson and Dumanski 1994). Smeck et al. (1983) concluded that numerical classifications have conceptual value, but none are totally acceptable. Nonetheless, numerical classification methods now play an increasingly important role in soil classification and mapping research.

Some Definitions of Soil Classification

Classification is the process of sorting or arranging of objects into groups on the basis of one or more objectives and according to a system or set of principles. *Categorization*, according to usage by cognitive scientists (Rosch and Lloyd 1978), involves the prior establishment of classes or groups of objects possessing one or more commonalities of properties and the placing of objects into preconceived and preformed groups according to quantitatively measurable criteria for category membership. This is done in a fashion that will provide maximum information with a minimum of classification effort while best representing the natural structure in the perceived world. Much of what we do in placing soils in groups is categorization rather than classification, but we bow to common usage and will use the term "classification" to cover both classification and categorization.

Class is a group of individuals or other units similar in selected properties and distinguished from all other classes by differences in these properties.

Taxon (pl. *taxa*) is a class of a formal classification system at any level of generalization or abstraction.

Category is a series or array of taxa produced by differentiation within the population being categorized at a given stated level of generalization. Note that we often use category and class as synonymous in our everyday conversation, but in the parlance of classification and categorization, a category is composed of an array or set of classes. Stated in another way, a category means a number of objects that are considered equivalent (Rosch 1978).

A *differentiating characteristic* is a property chosen as the basis for categorizing individuals (or classes of lower levels of generalization) into groups. There may be one or more characteristics used to represent or differentiate a class.

A *multiple-category* or *multicategorical system* is a hierarchical system of categories designed to classify a large and complex population that has a vertical structure of categories usually proceeding from more general, broadly defined categories to more specific, or narrowly defined, categories.

Taxonomy is a particular formalized system of classification or categorization developed for a specific purpose and categorized according to a set of prescribed differentiating characteristics. The term may also refer to the science of classification and categorization. In cognitive science, "taxonomy" is defined (Rosch et al. 1976) as a system by which categories are related to one another by means of class inclusion, with each category within the taxonomy entirely included within another, more generalized (higher) category, unless it is the highest category.

General Perspective of Soil Classification

People have a natural tendency, urge, and need to sort and classify the natural objects of their environment. Folk, or local, classifications (for example, those made by indigenous peoples) are based on recognition of natural breaks in readily perceivable characteristics. These classification systems tend to classify all natural objects in a local geographic area and make class distinctions according to technology locally available at a basic level of generalization roughly comparable to the level of the genus (Berlin 1978; Rosch 1978). Soils are no exception. Folk classification and names of soils have carried over into present-day classifications (Simonson 1985). Soils are present throughout the world, and many classification systems exist. Many soil classification systems that include only portions of the world are well constructed, but are limited in the worldwide transfer of information among soil scientists and others that seek information regarding soil properties.

The classification system used in a scientific discipline reflects the state of the art and knowledge in the field. Kubiena (1948) claims, “Show me your [classification] system and I will tell you how far you have come in the perceptions of your research problems.” The renowned physicist Ampére is reported to have said, “Perfect scientific classification is first possible when one knows everything concerning the classified natural objects” (Kubiena 1948).

Classification and categorization of natural phenomena (like soils) are generally done for one or more of the following purposes:

1. Organize knowledge of the subject.
2. Provide maximum knowledge about the subject with the least cognitive effort (perception, recall, and memory of properties).
3. Provide a map or organization chart of structure of the world we perceive and live in, to satisfy our natural curiosity, and for ease in communication.
4. Reveal and understand relationships among individuals and classes of the population of interest.
5. Identify and learn new relationships and principles not previously perceived in the population of interest.
6. Provide objects or classes as subjects for research and experimentation and/or research design.
7. Establish groups or subdivisions of the object classified and under study in a useful manner, such as:
   a) predicting behavior,
   b) identifying best uses,
   c) estimating their productivity,
   d) identifying potential problems and as basis for taking action to meet potential problems,
   e) facilitating easier transfer of information and technology.

Kinds of Soil Classification. We differentiate between natural or scientific classification and technical classification as follows.

Technical classifications of soils organize and classify objects into groups for specific applied purposes Cline (1949b). Technical classifications are often created to facilitate implementation of land use regulations or laws. Technical classifications utilize criteria specific to purposes of concern for a specific technology and ignore other soil properties. Technical classifications therefore rely on the state of the art and understanding of the intended use of the soil or the present land-use regulation. Technical classifications are subject to change as regulations change or the knowledge of the technology or land use practice changes. These systems are seldom useful in presenting the state of the art in soil science, but serve a vital purpose in communicating information regarding specific soil properties to a variety of disciplines.

Natural or scientific classifications bring out relationships among the most important properties of the population being classified, without reference to any single specified applied or utilitarian purpose (Cline 1949b). In the field of logic, a natural classification is one that considers all the attributes of all the individuals in a population (Mill 1925). Those attributes having the greatest number of covariant or associated characteristics are selected as criteria in defining and separating various classes. From the standpoint of cognitive science, a natural classification is one in which a basic level of categorization is established within which as many attributes as can be identified and measured are common to all members of the category. In *Soil Taxonomy* this basic level unit would be roughly equivalent to the soil series. A superordinate level is also established. This level is organized around a few prototype members that display a limited number of attributes that have been selected as having the greatest number of covariant or associate characteristics. In *Soil Taxonomy,* this level is the soil order. Intermediate levels between basic and superordinate levels are established in which an increasing number of attributes are shared by all or most members (Rosch et al. 1976).

Principles of Soil Classification

Now we need to consider some principles as background for our discussion of soil classification systems. Some aspects are unique to pedology, but most are general principles germane to the natural classification of all natural objects.

Genetic-Thread Principle. The theories and concepts of soil genesis provide a framework for determining the significance and relevance of soil properties for use as differentiating characteristics. This principle parallels the use of theories of evolution as a framework for taxonomy in the plant and animal kingdoms.

Principle of Accumulating Differentia. In a multiple-category classification system, differentiating characteristics accumulate or pyramid from the higher levels of generalization to the lower levels of the system. As a result, classes at the lower levels are defined and differentiated not only by the differentiating characteristic(s) used at a given categorical level but also by those used as differentia at the higher levels. In the lowest category, a large number of differentiating characteristics have been accumulated so that the classes are quite narrowly defined.

Principle of Wholeness of Taxonomic Categories. All individuals of the population must be classified in each category, according to the characteristics selected as differentiating at that level. Another way of stating this is that any differentiating characteristic should classify all the individuals of a given population. Some soil classification systems have violated this principle by omission of certain kinds of soils from classification at one or more of the categorical levels.

Ceiling of Independence Principle. A property or characteristic used as a differentiating characteristic in a higher category must not separate individuals grouped together in a lower category. Every characteristic used to define individuals in a category is limited in its use by a "ceiling" categorical level above which it cannot be applied without injecting confusing and inappropriate cleavages in lower categories.

Succession of Classifications

Classification undergoes revision as the body of knowledge on which it is based expands. In this sense, classification succession is a phenomenon common to all disciplines. We should always keep in mind that classification systems are developed by human beings to organize ideas and properties in the most useful manner (Cline 1963). Classification systems are abstracts of "knowledge" and of concepts derived from knowledge, based on the past experiences, available data, and present biases of people (Cline 1961). The data we consider to be "facts" are facts only within the context and perspective of the operations by which these data were obtained (Bridgman 1927). An example in the field of soil science is the previously held concept, and hence "fact," that exchangeable hydrogen was the main source of soil acidity in acidic mineral soils, whereas it is now generally accepted that exchangeable aluminum is the main contributor to soil acidity in such soils (Coleman et al. 1959; Jenny 1961c). A comparison of the general basis and class definitions for a soil classification system published in the 1938 USDA *Yearbook of Agriculture* (Baldwin et al. 1938) with those in *Soil Taxonomy* (Soil Survey Staff 1975, 1999) is a more complete and complex example of chronological secession.

In the development of the total field of science, we can see many examples of the uncovering of new facts requiring rather complete reorganization of the theories and laws making up the body of understanding in a particular field and thereby requiring extensive changes in the classifications based on this understanding (Bridgman 1927; Cline 1961). The point is that we must accept the provisional, ephemeral, changing state of current knowledge and consequently of classifications based on that knowledge, particularly in soil science, where not only methods of analysis are evolving, but intensity of observations are not uniform throughout the world. We must be prepared to accept additional changes in soil classification as soil science research continues; indeed, we should help make them.

Avoiding Rigor Mortis in Classification

It is easy for a classification system to prejudice the future and for us to become prisoners of our own taxonomy (Cline 1961). At times and in some places this has been a particular problem with soil classification systems. The nomenclature of classification systems is preserved in the written text and fixed conventional wisdom; incomplete data and tentative hypotheses harden into dogma, thereby restricting acceptance of new ideas, concepts, and research patterns needed for acquisition of new facts. Therefore, a classification system, particularly in a field such as soil science, must have a self-destruct mechanism—a procedure for continuing reevaluation of the body of theories making up the genetic thread in our taxonomy. Also, we must avoid selection of soil genesis theories and hypotheses as basic differentiating characteristics, although we may use them, with caution, as guides to relevant soil properties that may be used as differentia.

Historic Perspective of Soil Classification

The evolution of soil classification can be subdivided into five general periods: (1) an early technical era; (2) the founding period of pedology by the Russian group of soil scientists; (3) the early American period; (4) the middle period of general development of soil classification and soil surveys in the world, and especially in the United States, which we refer to as the Marbut period; and (5) the present modern period of quantitative taxonomy.

Early Technical Period. Soil classification had its inception and flowering in Western Europe in the middle and latter parts of the nineteenth century. For example, Thaer (1853) published a classification that combined textural (particle-size distribution) properties as a primary higher category, with agricultural suitability and productivity as classes of a lower category. He identified six kinds of soil: clay, loam, sandy loam, loamy sand, sand, and humus. As an example of classes within these six kinds of soil, he recognized four classes within the group of clay soils: black klei-heavy wheat soil, strong wheat soil, weak wheat soil, and thin wheat soil.

Fallou (1862) devised the following soil classification largely based on geologic origin and lithologic composition of what we now call parent material:

Class 1: Residual Soils	Class 2: Alluvial Soils
Soils of limestone rocks	Gravel soils
Soils of feldspathic rocks	Marl soils
Soils of clay rocks	Loam soils
Soils of quartz-bearing rocks	Moor soils

Von Richthofen (1886) worked out a system of soil classification with a strong geologic basis and nomenclature similar to Fallou's system:

A. Residual soil types
   1. Disintegrated rock
   2. Deeply weathered rock
   3. Eluvial soils of plateaus
   4. Colluvial loam
   5. Laterite
   6. Organic soils: humus, moor, peat
   7. Undissolved residues

B. Accumulated soil types
   1. Coarse sediments of continental waters
   2. Fine-grained sediments of continental waters
   3. Chemical deposits in fresh waters
   4. Marine soils
   5. Glacial deposits
   6. Volcanic ash
   7. Eolian accumulations

These systems of classification used differentiating characteristics of geologic deposits, not properties of the soils. We have presented them for comparison with more recent and comprehensive classification systems and a historical record of the first stirrings of a new field: soil science.

The Founding of Pedology. The founding of pedology, or soil science as an independent science, can be traced to soil studies on the Moscow Plains in Russia. On the great Central Russian upland, uniform loess-like parent materials extend for hundreds of miles. An increasing temperature gradient is imposed on this area from north to south, and decreasing annual rainfall and moisture gradient extends from west to east. Associated with these climatic parameters are important vegetation pattern differences, especially the major shift from forest to grassland steppes. These factors of climate and vegetation left their imprint on the relatively uniform parent material, producing distinct soil differences. These differences were noted by the founder of modern pedology, V. V. Dokuchaev, who first understood the significance of soil differences not related to geologic material and thereby established the concept of soil as an independent natural body.

V. V. Dokuchaev was born in a middle-class family in Smolensk, where he graduated from a seminary. Upon entering Petersburg University, he received training as a geologist. He did his first fieldwork in the Smolensk area and prepared his first published work, "Origin of River Valleys." About this time he met the Russian statistician and natural scientist, Chevlovsky, who had prepared the first soil map of Russia, using folk definitions and names (e.g., "poor" and "rich" soils). As a result Dokuchaev became particularly interested in the "Chernozem," the "richest" of soils in Russia. He received a grant from the Free Economic Society in St. Petersburg for studies of these soils (apparently the first grant for scientific research on soils). In 1883 he published his classic monograph, *Russian Chernozems* (Dokuchaev 1883). This was the first published work on soil as a natural body formed by action of a set of soil-forming factors producing genetic layers in the parent material. He followed this monumental monograph with a series of publications on soil genesis and classification, including the first publication with a classification of soils based on the properties and soil-forming factors of the soils themselves (Dokuchaev 1886). Dokuchaev was concerned about and interested in not only the technical and scientific aspects of soil classification but also in practical applications of soil classification. For example, he interpreted his classification of soils of the Nizhnii-Novgorod (now Gorki) region in terms of tax value assessment (Dokuchaev 1886). Also, he became concerned about very serious droughts in parts of the Russian steppes and established a series of experimental plots on different soils. From the results obtained, he recommended forest belts and other procedures for wind protection and water conservation.

Dokuchaev has been erroneously regarded and described by certain latter-day soil scientists as a theoretical scientist who related soil formation and classification solely to climates, for example, a "climatic soil scientist." Although much of Dokuchaev's writing emphasized the importance of climate, this misconception apparently stems in part from a statement made by Glinka (1931) in a widely translated book in which he listed five main soil types as proposed by Dokuchaev, linked to climate as the sole or main soil former. However, we should point out that a sentence translated from Dokuchaev's classic work, *Russian Chernozems*, states that "soils must be classified and studied according to their profiles."

Dokuchaev is reported as being a very sociable man who liked people. His writings and dynamic personality attracted many capable students. Those of his students most closely involved in soil classification were N. M. Sibirtsev and K. D. Glinka. Their contributions also stand as significant monuments in the Russian pedological founding period.

Sibirtsev (1860–1899) developed the concept of soil zones, a powerful idea that certain kinds of soils are associated with certain "climate-vegetational" or ecological regions. This concept is a basic part of many soil classification systems. Sibirtsev completed his classic text, *Soil Science* (*Pochvovedenie*) in 1901, apparently the first text on soils and soil classification, but unfortunately he met an untimely death from tuberculosis.

Glinka (1867–1927) was the most influential and prolific writer among Dokuchaev's pupils and the best known in the Western world because of widespread translations of his works: *The Types of Soil Formation, Their Classification and Geographical Distribution* (1914), *The Great Soil Groups of the World* (1927), and his classic *Treatise on Soil Science* (1931). His first book introduced the new Russian concepts of soils, soil classification, and the major soil type names of Chernozem, Podzol, and Solonetz to the Western world. Glinka emphasized soil geography, soil formation, and weathering processes. He was a brilliant lecturer and organizer, being responsible for the organization of soil science in Russia. A number of other outstanding Russian soil scientists were active and prominent in the latter part of the Russian pedologic founding period, but we discuss only the pioneers of this particular period.

The Early American Period. Ruffin (1832) pointed out the need for a soil classification program in the United States. E. W. Hilgard pioneered early soil classifications and mapping in the United States. He was a geologist for the state of Mississippi and published a classic pioneer work on soils of Mississippi and later was responsible for establishment of work on soil science in California, especially with respect to sodic and saline soils. He apparently was first in America to conceive of soil as a natural body and pointed out correlations between soil properties on the one hand and vegetation and climate as causal factors. It has been suggested that Dokuchaev was a follower of

Hilgard. Although Dokuchaev's work came a little later, there is no clear indication that the two were in contact or knew of each other's work. Hilgard's concepts and ideas about soil and soil genesis were not applied in operational soil surveys in America, and introduction of similar ideas was not to come until more than 50 years later.

In 1894 a division was created within the Weather Bureau for "the study of climatology in its relation to soils," and in 1901 a Bureau of Soils was created in the USDA (Whitney 1905). Emphasis was placed on technical or single-factor classifications of soil in operational programs of soil surveys in the United States. In these programs, there was bias toward geologic techniques and nomenclature, though there were some notable exceptions. Milton Whitney developed the first American soil classification system related to soil survey and used as a basis for soil-mapping operations. This system was published in 1909 (Whitney 1909), although actual soil surveys started in the United States about 1899 (U.S. Department of Agriculture 1899). Whitney's system was a broad classification, mainly according to physiographic regions or provinces and the texture (particle-size distribution) of the soil. Whitney and associates established as the highest taxonomic category the soil province, composed of soils within the same physiographic regions, such as coastal plains or piedmont. Soils within a province formed from similar geologic materials (such as marine sediments or glacial till) were defined as a "series," a term still in use today in the United States for the taxa of the lowest taxonomic category. Soils were subdivided within the series according to texture to form the lowest taxonomic category and the mapping unit: the soil type. Whitney and his associates defined "texture" more broadly than is now the case in the United States. "Texture" included particle-size distribution, soil consistence, organic matter content, aggregation, and related properties. This system was widely used as a basis for soil surveys in many parts of the United States.

Coffey (1912) was apparently the first in the United States to propose soils as independent natural bodies that should be classified on the basis of their own properties and that differences in these properties were due to climatic and associated "vegetational" differences from place to place. He proposed five great soil groups (apparently the first use of this term, which became widely used): arid, dark-colored prairie, light-colored timbered, black swamp, and organic. However, his concepts and proposals apparently were not generally accepted and were not made the basis for any operational soil survey program, but his ideas did serve as a forerunner and signaled changes to come in soil classification in the United States.

The Middle American Period. C. F. Marbut was the central figure in the evolution of soil classification and survey during this period in the United States. Born in 1863 and raised on a Missouri farm, he was trained in geology, especially geomorphology. He undertook graduate study in that field at Harvard University under the well-known American geomorphologist William Morris Davis. He became interested in soils and joined the USDA Bureau of Soils in 1910. He introduced the concepts of Dokuchaev, Glinka, and Sibirtsev to the United States after translating a German edition of Glinka's work on types of soil formation and soil groups of the world (Glinka 1914) into English. This introduced the soil-forming factors of climate and vegetation and

Table 1.1. Soil classification by Marbut (1935)

Category 6	Pedalfers	Pedocals
Category 5	Soils from mechanically comminuted materials	Soils from mechanically comminuted materials
	Soils from siallitic decomposition products	
	Soils from allitic decomposition products	
Category 4	Tundra	Chernozems
	Podzols	Dark brown soils
	Gray-brown Podzolic	Brown soils
	Red soils	Gray soils
	Yellow soils	Pedocalic soils of Arctic and tropical regions
	Lateritic soils	
	Laterite soils	
Category 3	Groups of mature but related soil series	Groups of mature but related soil series
	Swamp soils	Swamp soils
	Glei soils	Glei soils
	Rendzina	Rendzina
	Alluvial soils	Alluvial soils
	Immature soils on slopes	Immature soils on slopes
	Salty soils	Salty soils
	Alkali soils	Alkali soils
	Peat soils	Peat soils
Category 2	Soil series	Soil series
Category 1	Soil units or types	Soil units or types

reduced the emphasis on the geologic nature and origin of soil materials as developed by Whitney. His ideas on classification evolved in successive steps (Marbut 1922, 1928a, 1928b) and culminated in his masterwork on soil classification published in the *Atlas of American Agriculture* (Marbut 1935). A summary outline of this 1935 classification is presented in Table 1.1.

This dedicated man, who died in 1935 while in China, must be considered to be the founder of pedology in the United States, based on his many contributions in addition to his worldwide influence (Krusekopf 1942). Some of his many contributions were:

1. Establishment of the soil profile as the fundamental unit of study. He focused attention on properties of soils themselves rather than their geologic relationships or broad soil-forming factors.
2. Preparation of the first truly multiple-category system of soil classification.
3. Establishment of criteria for identifying and naming soil series.

Although Marbut's classification system advanced soil science, additional information and further developments and improvements identified some difficulties and problems. Some of these are as follows:

1. His multiple category system was not truly comprehensive. The emphasis on "normal" soils on "normal landscapes" (meaning the well-drained soils on hillslopes)

Table 1.2. Soil classification in the 1938 USDA *Yearbook of Agriculture* (highest two categories only)

Category 6 Order		Category 5 Suborder
		Soils of the cold zone
	Pedocals	1. Light-colored soils of arid regions
		2. Dark-colored soils of the semiarid, arid, subhumid, and humid grasslands
Zonal soils		3. Soils of forest-grassland transition
	Pedalfers	4. Light-colored podzolized soils of the timbered regions
		5. Lateritic soils of forested warm-temperate and tropical regions
Intrazonal soils		1. Halomorphic (saline and alkali) soils of imperfectly drained arid regions and littoral deposits
		2. Hydromorphic soils of marshes, swamps, seep areas, and flats
		3. Calomorphic
Azonal soils		No suborders

omitted the classification of "immature and abnormal" soils in one or more categories, thus violating the principle of wholeness of taxonomic categories.

2. Certain criteria for differentiation (differentiating characteristics) based on assumed genesis or genetic interferences were incomplete or incorrect. For example, the assumption that Zonal ("normal") soils could be divided into two broad classes—one in which calcium carbonate accumulates (Pedocals) and another in which aluminum and iron accumulate (Pedalfers)—has been shown to be inadequate and not satisfactory. A soil that is a "normal" Pedocal in one region was considered an Intrazonal (not normal) Pedocal in another region due to slight differences in parent material or landscape position. Some soils accumulate both $CaCO_3$ and Al-Fe compounds. For these and related reasons, this differentiating characteristic and these particular classes had to be abandoned.
3. His "normal soil on a normal landscape" concept as a basic frame of reference for soil classification was not appropriate because it ignored more poorly drained soils in complex landforms. Further, differences in soil age, as well as differences in climate over time, made it difficult, if not impossible, to establish which soil is the "normal" soil of reference in many landscapes.
4. Extreme emphasis was given the two-dimensional soil profile, largely ignoring the three-dimensional aspects of soils on the landscape.

As new information was obtained and evolution of concepts took place, a comprehensive effort was made to improve Marbut's classification of all known soils in the United States. This revision was published in the 1938 USDA *Yearbook of Agriculture, Soils and Men* (Baldwin et al. 1938). An outline of the orders and suborders proposed by Baldwin and coworkers is presented in Table 1.2.

The Modern Quantitative Period. After World War II, efforts were undertaken to revise the 1938 USDA *Yearbook* classification (Thorp and Smith 1949; Riecken and Smith 1949). In these revisions, new Great Soil Groups were added, and definitions were revised and sharpened, but the authors determined a need for a new more quantitative approach to soil classification. In 1951, a decision was made to develop a new soil classification system in the United States (Smith 1968).

Reasons for undertaking development of a new system of classification in the United States are summarized in the following points, drawn in part from Kellogg (1963a), Simonson (1952a, 1952b, 1986a, 1986b, 1986c), and Smith (1968, 1983, 1986):

1. The highest category of the 1938 system, based on zonality, did not provide mutually exclusive taxa; it was not possible to define clearly the differences between Zonal and Intrazonal soils.
2. Classification at higher levels of the 1938 system, as well as those of other existing systems, was based on external environmental factors and assumed genesis as differentiating characteristics, not properties of soils themselves. Thus, there was risk of "prejudicing the future" by incorporating assumed soil genesis concepts, and it was difficult to classify certain soils because of uncertainties or disagreement among soil scientists concerning their genesis.
3. Some definitions of taxa were based on virgin soil profiles under their native vegetation, without allowing for modifications due to tillage and/or erosion.
4. Too much emphasis had been placed on soil color as a differentiating characteristic without consideration of relevance or number of accessory characteristics associated with color.
5. Many of the lower categorical taxa levels were defined in terms of comparative and subjective definitions. Quantitative, objective differentiating characteristics of taxa were needed to assure interpersonal agreement on classification and to make taxa suitable objects for research and interpretations.
6. Some soils were not identified in all categorical levels. Provisions were needed for all known soils to be classed at all categorical levels, and flexibility was needed to accommodate classification of newly discovered soils in sparsely surveyed areas of the United States and developing areas of the world.
7. Soil families had not been clearly defined as a category, and taxa within this category were poorly identified.
8. The nomenclature used in existing systems was a collection from several sources, both folk names in several languages and coined names. Often the same term held different meaning for different people, translation was difficult, and the naming of soil intergrades was difficult if not impossible.

As a consequence, development of a completely new system, above the levels of the soil series, was started within the USDA under the leadership of G. D. Smith. Smith was a tireless worker whose excellent scientific skills, practical judgments, and friendly personality enlisted the cooperation of soil scientists in the United States universities and many overseas scientists in the task of developing *Soil Taxonomy*.

The development of this new comprehensive system was by a series of *Approximations*, each widely circulated for criticism and comment. *Soil Classification, A Comprehensive System–7th Approximation* was published and distributed at the International Soil Science Society meeting in 1960 (Soil Survey Staff 1960) to ensure worldwide circulation and, hence, a broader spectrum of comment and criticism. Both comments received and further studies were used as a basis for supplements published in 1964 and 1967 (Soil Survey Staff 1964, 1967). After 15 years of intense efforts to test and document criteria, *Soil Taxonomy* was published (Soil Survey Staff 1975). From 1975 to 1998 the *Keys to Soil Taxonomy* underwent seven published revisions. In 1999 the second edition of *Soil Taxonomy* was published (Soil Survey Staff 1999). We use this edition, plus the most recent edition of the *Keys to Soil Taxonomy* (Soil Survey Staff 2010), as a basis for presentation in this book. (See Chapter 7.)

Soil Morphology

This aspect of soil science deals with the description of a soil at a specific site. Methods of documenting soil morphology have advanced greatly since the conception and initiation of field soil surveys in the late nineteenth century. Early soil descriptions were highly personalized descriptions that lacked quantification and standardization needed for effective communication. For example, early descriptions of soil color included such picturesque terms as *lemon yellow*, *coffee brown,* and *mouse gray*. The use of the Munsell color system now quantifies the color measurement and provides standard terminology.

In this section we introduce some basic definitions and concepts of soil morphology. A more detailed and expanded discussion of morphology is presented in Chapter 2. Details of the methods of field descriptions and recording field observations are given in the *Soil Survey Manual* (Soil Survey Staff 1951, 1962, 1993) and in the *Field Book for Describing and Sampling Soils* (Schoeneberger et al. 2002).

Soil is that portion of the earth's surface located in the vertical space between air above and geologic substratum below, with horizontal limits bounded by materials such as deep water, ice, and rock outcrop that are not considered soil (Figure 1.1). At any location a vertical exposure of the soil is known as a soil profile.

A complete *soil profile* is a vertical exposure that includes all material that has been altered by chemical, physical, and biological reactions that are caused entirely or in part because of proximity to the land surface (Figure 1.1). A soil profile may be observed in a freshly dug pit, along a road bank, or in many other places. In practice it is not always possible to expose a complete soil profile. It is often difficult to determine at what depth the material ceases to exhibit properties resulting from reactions related to its position near the land surface. That is to say, "Where does the soil stop and the 'nonsoil' begin?" Examination and description to a depth of 2 m or more may be necessary to adequately examine a soil to interpret soil-forming processes. By convention, *Soil Taxonomy* restricts soil classification to soil features that occur within 2 m of the soil surface. If a hard root-restricting layer is encountered at a shallower depth, every effort available should be made to identify the restricting layer. The soil profile is the basic unit for observing soil, but several other units of soil related to a soil profile need to be introduced.

The *soil solum* is an incomplete soil profile (Figure 1.1) that may be simply defined as "the genetic soil developed by soil-building forces" (Soil Survey Staff 1975). In this context, the solum includes O, A, E, and B horizons, defined in Chapter 2. Although this definition seems simple enough, much confusion can arise when application is made in the field. Determination of the lower boundary of a B horizon, hence, the solum may be difficult (Chizhikov 1968). The maximum rooting depth of adapted vegetation is another criterion for identifying the lower limit of the solum. The primary difference between soil and geologic material is the presence of living plant roots, deposits of organic materials originating from vegetation, morphologic evidence of root penetration, or evidence of bioturbation by soil fauna. It can be argued that any portion of the earth's crust that is reached by plant roots has been changed from geologic material into soil. In this context, it can be reasoned that the *solum* includes horizons significantly affected by soil-forming processes (that is, O, A, E, and B horizons), but that the *soil* extends to a greater depth, the depth affected by plant roots. Thus, a complete soil description does not stop at the bottom of the solum, but includes one or more underlying layers, ideally to a depth of 2 m.

The *control section* (Figure 1.1) is a portion of the soil profile delimited in terms of specified depths in family category of *Soil Taxonomy* (Chapter 7), primarily for identifying particle-size and mineralogy classes. The location of the control section differs depending on a number of soil properties and classification criteria. The control section of the soil profile has utility in soil classification because it is little influenced by common management practices such as plowing and fertilization, and thus provides a volume of soil where soil chemical and physical properties remain relatively constant over time. This arbitrary working rule, considered necessary to aid uniform soil classification, is not a conceptual aid in a study of soil genesis.

Soil Individuals on the Landscape

Although the soil profile is the basic unit for observing the vertical arrangement of soil components, a soil profile provides only one observational point of a given kind of soil. Like other natural bodies, an individual soil body is bounded laterally by other soil bodies or by nonsoil material. For example, in a forest, an individual tree of one species may be surrounded by other species of trees. While the lateral margins of individual trees can be easily seen, individual soils merge with adjacent soil bodies, and it becomes necessary to establish conventions by which individual kinds of soil can be differentiated. Practical considerations limit the spatial detail useful in delineating the minimum volume of observation required in establishing soil individuals.

A *pedon* is the smallest volume of soil that should be recognized as a soil individual (Figure 1.1). The pedon has been defined as follows (Soil Survey Staff 1999, p. 11):

> A pedon has the smallest volume for which one should describe and sample the soil to represent the nature and arrangement of its horizons and variability in the properties that are preserved in samples. A pedon is comparable in some ways to the unit cell of a crystal. It has three dimensions. Its lower limit is the somewhat vague limit between the soil and "nonsoil" below. Its lateral dimensions are large enough to represent the nature of any horizons and variability that may be present. A horizon may vary in thickness or in composition, or may be discontinuous. The minimal horizontal area of a pedon is arbitrarily set as 1 m^2, but ranges to 10 m^2, depending upon the variability in the soil.

The pedon concept enlarges upon that of the soil profile to include both lateral and vertical dimensions of a soil and puts limits on the volume to be considered as an individual soil. In essence a pedon includes properties that repeat within a lateral distance of 10 meters or less. Judgment must be used in application of the pedon definition in field situations. It is not prudent to describe and sample the material in an animal burrow or the auger hole made by a previous soil scientist that may vertically traverse a soil, unless one is researching the composition of material in animal burrows or auger holes. Narrow cracks in bedrock underlying the solum may be penetrated by roots and should be described as a feature of that bedrock layer. Sampling may not be possible.

Although a pedon is the smallest volume that can be considered a soil individual, it is seldom the case that a soil individual can be adequately described and defined from one pedon. The range of properties observed in several similar pedons, a *polypedon* (Figure 1.1), are aggregated to define a soil individual.

A *soil individual* is a defined range of soil properties used to classify soils. Judgment is required in defining a soil individual. Soil individuals are defined by experienced soil scientists with due consideration to the present and potential uses. The defined range of pedon properties that identify an individual soil must mutually interface with other soil individuals so as to include all soils. An individual soil is the smallest aggregate of defined soil properties used to identify taxonomic units in constructing soil taxonomic systems. A soil individual is a defined concept, with a range of properties seldom, if ever, totally present in one pedon. The soil individuals used as taxa in *Soil Taxonomy* (Soil Survey Staff 1999) are known as series. (See Chapter 7.)

The task of establishing consistently recognizable limits for soil individuals is daunting. In the absence of a specific characteristic that is present and of significance in all soils, like DNA in humans, a variety of quantitative characteristics are necessary to identify soil individuals. This results in definitions that often seem complicated by the specific conditions they establish. However, vaguely defined criteria or criteria that can be recognized only under very special conditions add confusion and distrust of the system. Although soil property associations with vegetation, topography, and parent material can indicate where soil differences are likely to occur, such associations should not be used as criteria in defining soil individuals because such associations are not universal. Criteria that are dependent on management practices are usually too transient to be useful. A change in soil management practices can render meaningless soil classes based on conditions detectable under the old practices. Only properties that can be consistently identified and measured within the soil under a range of soil management practices should be used as criteria for establishing taxa.

Precisely defining individual soil taxa is only the first step in conveying soil information to most people who seek information about soils. A pedon or an individual soil has little practical value unless it is used to identify an area of land. Rarely do all the soils spatially clustered on the landscape belong to the same taxon. The spatial intermingling of taxonomically different individuals is common among all natural entities. For example, in a lawn that contains one predominant species of grass, we often see a different plant species growing and call it a weed. Taxonomically different soil individuals are frequently "weeds" in an area where the soils are mostly of one taxon, but the identity of that "soil weed" only becomes apparent when we examine several adjacent profiles or pedons.

Soil map units are not part of the classification system of *Soil Taxonomy*, but are constructed during the course of a soil survey to identify populations of taxonomically identified soil individuals within a spatial area on the land. Soil mapping is the action, or field, phase of the genesis-classification-morphology-survey-characterization-interpretation continuum. Butler (1980) describes the soil survey as "one of the basic technologies of soil science." Soil maps are cartographic representations depicting the location of a population of soil individuals (Fridland 1976a, 1976b). A taxon name is commonly used as part of the map unit name, but unavoidably soil map units contain soils not taxonomically defined by that name (inclusions, "soil weeds"), confusing the nonsoil scientist (Cline 1977b). Readers are cautioned to remember that when used to name a map unit, the name of a soil individual, usually a soil series name, almost always identifies a spatial association of several soils. This subject is discussed in greater detail in Chapter 20.

Other Soil Genesis and Classification Activities

Soil characterization is the measurement of soil properties by laboratory procedures, using soil samples from pedons, the morphology of which has been described by standard procedures and nomenclature. Soil characterization is conducted to classify soils, to determine chemical and physical properties not visible in field examination, and to gain a better understanding of soil genesis. Laboratory procedures for standard soil characterization conform to protocols to assure that direct comparisons can be made among the individual soils sampled. The Natural Resources Conservation Service of the USDA now has characterization data for more than 20,000 pedons from the United States and more than 1,100 pedons from other countries.

Soil interpretations are predictions of how a soil, or an area of soils within map unit delineations, responds to land use and management practices. It is through this process that the data, knowledge, estimates, hypotheses, and theories are put to the test of practical, applied uses. The purpose of soil classification and mapping activities in the United States has been to provide interpretations for technical assistance in soil conservation programs, for planning agricultural programs, and as a basis for financing credit (Kellogg 1960; Smith 1968). Since the first soil survey reports and maps were produced in the United States, technologies of soil use have changed, necessitating many changes and additions in the soil interpretations. Interpretive rating guides for more than 70 specific types of interpretations are presented in the *National Soil Survey Handbook* (Soil Survey Staff 2009), a constantly updated handbook used in the writing of soil survey publications.

Soil Survey

Soil survey encompasses mapping, classification, characterization, and interpretations of soils. The field mapping of soils by many individual soil scientists requires quality control activities, commonly called correlation, to assure compatibility and comparability among soil surveys per agreed-upon principles, nomenclature, and established classifications. Soil characterization is an important aspect of quality control. Interpretations of soil taxa need to be uniform, but interpretations of mapping units also must reflect local conditions and spatial associations unique to individual soil survey areas. The field observations are compiled and recorded on hard copy maps and in digital geographical information system (GIS) formats. While not directly undertaken for soil genesis research, soil survey provides the factual observations necessary to formulate and test pedogenic theories.

Systematic soil survey was formally organized and initiated in the United States in 1899 (Whitney 1900, 1905; Simonson 1986a). By 1905 the USDA Bureau of Soils had mapped 56.9 million acres (Whitney 1905). The earliest soil surveys adhered to a concept of soil as a geologic material with some admixture of organic matter (Simonson 1986a). The term "soil series" was derived from the concept of a series of soils found on a particular geologic formation. As such, a series was consistent with what is now known as a map unit, but now the term "series" is used to identify an individual soil or taxon. Most early soil mapping was at the scale of 1 inch per mile (1:63,360) and was accomplished with the aid of a plane table and alidade to prepare the base map as the field survey proceeded. In the 1930s the practice of using air photos as a base for the field mapping was introduced, soon followed by the practice of publishing the soil maps on an air photo base (Simonson 1986c). Present-day soil surveys use digital orthophoto imagery as the base. Most surveys are published digitally and are accessible at the Natural Resources Conservation Service Web Soil Survey site (http://websoilsurvey.nrcs.usda.gov/app/HomePage.htm).

Progress in concepts of soil and technology of soil mapping has proceeded in fits and starts. A number of states and other federal agencies undertook soil-mapping programs, leading to the formation of the National Cooperative Soil Survey in the 1950s. The Soil Conservation Service, renamed the Natural Resources Conservation Service in 1995, was responsible for correlating the surveys and publishing the reports and maps. The scale of mapping increased to show more soil pattern detail, especially in agricultural areas. Present soil map scales in agricultural areas range from 1:12,000 to 1:24,000 or higher. The number of soil series and erosion, slope, and other phases has increased algebraically in the past 70 years, and the use of soil maps for nonagricultural planning, inventory, engineering, and evaluation purposes has greatly increased in this period.

The *Soil Survey Manual* (Kellogg 1937; Soil Survey Staff 1951, 1962, 1993) has been periodically upgraded and extensively revised. Quality assurance and compatibility among surveys is provided through the use of the *Manual* and the constant updating of the *National Soils Handbook* (Soil Survey Staff 2009). The adoption of *Soil Taxonomy* has impacted the soil survey in the United States by providing precise and quantitative criteria for identifying and classifying soil characteristics. This has improved the quality of interpretive information in soil survey reports.

Although soil surveys have been conducted in the United States for more than 100 years and almost all of the country has been surveyed once and much of it two or more times, the demand for soil survey continues, and old surveys are periodically updated with new soils information. At least three rather universal observations can be advanced to explain the continued demand for information obtained in a soil survey.

First, the scientific understanding and technology used to measure soil properties continues to advance. Soil scientists are now able to obtain measurements of soil properties that could only be estimated or perhaps were unknown only a few years ago.

Second, technologies that use soil continually evolve. Agronomic and forestry interpretations need to consider new cultivars and cultivation practices. The function of different soils in waste disposal, heavy metal contamination, carbon sequestration, and wetland preservation are but a few of the more recent topics soil scientists are asked to evaluate.

Third, relates to the rather old cliché "they aren't making any more land." As more intense use is made of land, especially in urbanizing areas, there is a demand for more precise identification of small parcels of land. This requires that soil scientists map at scales much larger than in previous times and be able to provide interpretations of areas that were too small to be identified in older soil surveys.

Perspective

In this introductory chapter we have outlined the history, described and defined the main themes of this book: soil genesis and classification, sometimes referred to collectively as pedology. Soil genesis is sometimes referred to as soil formation and soil classification as soil taxonomy. We prefer the terms *soil genesis* and *classification* and reserve the term *Soil Taxonomy* (Soil Survey Staff 1999) for the comprehensive scientific classification system now used in many countries. Soil genesis deals with studies, concepts, theories, factors, and processes responsible for soil development and change. Classification deals with the categorization of soils into scientific or technical groupings, and the cognitive basis of this categorization. The identification and definitions of soil profile, pedon, soil individual and soil map units presented are fundamental to discussions that follow. Brief definitions, descriptions, and histories of the important allied and/or corollary activities of soil morphology, survey, interpretation, and characterization have also been presented.

Morphology and Composition of Soils

2

Soil morphology deals with the form, structure, and organization of the soil material. Soil morphology is ordinarily first observed, described, and studied in the field, but investigation can be continued in the laboratory with optical and electron microscopes. Field observations with the unaided eye or with a hand lens are considered macromorphology, whereas observations utilizing a microscope are considered micromorphology. Morphology describes and measures a wide range of soil properties, and includes as assessment of soil particles and aggregates that provide estimates of soil void characteristics and hydraulic properties (Lin et al. 1999a, 1999b).

Soil composition includes chemical, physical, and mineralogical measurements of soil material removed from defined positions (horizons or layers) within pedons. Soil moisture and temperature dynamics are important components of soils that vary from year to year but can be characterized and classified within ranges that permit identification of many potential soil uses.

New methods for analyzing soil are constantly being developed and tested. As stated in Chapter 1, soil classification follows science. A comprehensive soil classification system must provide a basis for comparing all soils and therefore must utilize comparable methods to determine soil composition. We confine discussions to the methods most commonly used to characterize soil for classification in *Soil Taxonomy*. Researchers should not be limited to the methods presented here, but only after an analytical technique has been successfully used on a large number of soils and a substantial amount of data collected can it be evaluated and utilized to classify soils.

Soil Macromorphology: Examination and Description

Macromorphology is best evaluated from the in situ examination of the soil profile. A recently dug pit large enough for observation of a pedon is desirable. Old exposures such as road cuts and ditches are acceptable only for preliminary examination because morphological features often become altered after prolonged exposure. The exposed profile should be probed by hand, with the aid of a knife or small pick to remove any alterations resulting from digging equipment and to expose the natural condition of the soil. Following this cleanup is a good time to photograph the profile, including a tape measure or some other suitable reference for scale. The profile examination

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

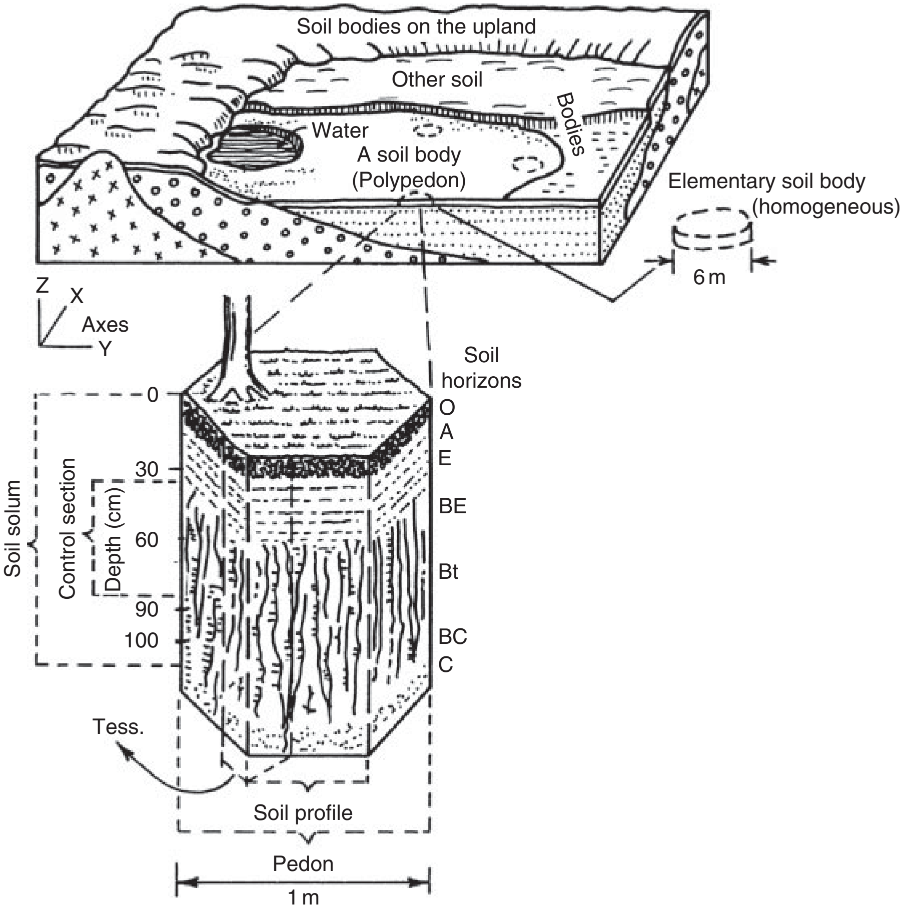

Figure 2.1. Profile, pedon, and solum. This pedon is from one of three elementary soil bodies that are inclusions in a large body.

begins by marking the depths in the profile where differences are observed in any number of soil properties including color, structure, texture, root distribution, rock fragment content, cementation, and/or carbonate content, as shown in Figure 2.1. This is the first approximation of soil horizon boundaries. The soil horizon boundaries may be adjusted as examination progresses, but it is desirable to establish an approximate boundary before beginning detailed examination of each horizon.

Recognizing Soil Horizons and Describing Their Properties in the Field. Soil horizons are described according to several properties including color, texture, consistence, structure, nodules or concretions, ped coatings, pH (by field method), carbonate content, root and pore distribution, and boundary characteristics, and horizon continuity. The following definitions of soil properties are based on the current practices in the United States (Soil Survey Division Staff 1993; Schoeneberger et al. 2002).

Soil Color. Soil color is probably the most obvious feature of the soil and is easily seen by laypersons. There are a number of methods used to describe color, but the standard method for soils is the Munsell system (Munsell 1912). The Munsell system identifies color based on three measurable variables: hue, value, and chroma. *Hue* is the dominant spectral color and is related to wavelengths of light. *Value* is a measure of degree of darkness or lightness of the color and is related to the total amount of light reflected. *Chroma* is a measure of the purity or strength of spectral color. These three variables have been combined into reference charts that cover the range of colors found in soils and bound in book format. In the Munsell soil color book, the various hues are arranged by page, one hue to a page. The units of value are arranged vertically, and the units of chroma are arranged horizontally on each page. Opposite each page of color chips is a page of color symbols and corresponding English names. An example of a color notation made for a soil horizon color is 10YR 6/3. The interpretation of the notation is 10YR (10 yellow-red) hue, a value of 6, and a chroma of 3. The proper name for this color is pale brown.

Soil color is moisture dependent, especially with respect to color value. The moisture status of the soil, moist or dry, is noted when color is described. Moist is the water content at which added water does not change the color; dry is air dry. In some soils the color of the ped exterior differs from the color of the ped interior, and the colors of both parts of the ped are described. In some cases, the color of crushed and smoothed soil material is described (e.g., spodic materials, defined later in this chapter), but prolonged rubbing of the soil material prior to measuring soil color should be avoided. The visual appearance of soil color is also affected by the source of light. Early morning or late evening light tends to make soil colors appear redder. Heavy overcast or deep shade under forest canopy makes measurement of soil color in the field difficult. If soil color must be determined in a laboratory or office after field sampling, the sample should be examined in skylight by a window or out-of-doors. Fluorescent lighting must be avoided.

Many soil horizons have intimately mixed color patterns. The dominant color of the horizon (the matrix color) is recorded and the *mottling* (from *motley: variegate in color*) is described in terms of quantity, size, and contrast of the other colors. The spatial pattern of mottling should be noted. *Quantity* is indicated as a percentage of the surface area examined that each color represents:

Few = less than 2% of the area
Common = 2 to 20% of the area
Many = 20% or more of the area

Size refers to the length of the mottle if the length is at least two times the width, and to the width of the mottle if the length is less than two times the width. Five size classes are used as follows:

Fine = less than 2 mm
Medium = 2 to 5 mm

Coarse = 5 to 20 mm
Very Coarse = 20 to 76 mm
Extremely Coarse = 76 mm or more

Contrast is a notation of the visual distinction between the colors described as combinations of differences in hue, value, and chroma of the matrix and the mottle. (See Schoeneberger et al. 2002 for details.)

Faint = indistinct contrast that is evident and recognizable only upon close examination
Distinct = color differences between the matrix and the mottle are more apparent
Prominent = mottles are obvious and mottling is one of the outstanding features of the horizon

Standard practice is to list the matrix color first, followed by a description of the quantity, size, and contrast of the other colors in the mottled pattern. In some cases, the soil colors are so intricately mixed that a dominant matrix color is not obvious. Under these circumstances, the proportion of each of the colors is estimated and each is described using Munsell notation. Careful observation and notation of the spatial arrangement of the various red, red-yellow, and gray colors (redoximorphic features) with respect to structural aggregates, root channels, or other large voids are useful in interpreting seasonal patterns of saturation and reduction in the pedon (Vepraskas 1992).

Soil Texture. Soil texture is defined as “the relative proportions of the various soil separates in a soil material” (Soil Science Society of America 1996). Percentages of sand, silt, and clay are reported on a mass basis. The continuum of soil texture has been divided into 12 textural classes (Figure 2.2A) used to describe soil horizons. The textural classes sand, loamy sand, and sandy loam are subdivided into additional classes based on the proportions of the various classes of sand-sized particles (e.g., very fine sand, loamy coarse sand, fine sandy loam). The lower part of Figure 2.2A shows how solid mineral particles are grouped by size in three systems. For most soil purposes the USDA scale is used. The family particle size groups of *Soil Taxonomy* are shown in Figure 2.2B for comparison but are not used when describing soil profiles. The textural groups diagramed in Figure 2.2A are determined using only the mineral particles less than 2 mm in diameter. Wetting a sample, “working” it between the thumb and fingers and comparing the “feel” to samples of known texture is an art developed by soil scientists to estimate texture in the field.

A rock fragment modifier prefaces the textural name if particles larger than 2 mm compose 15% or more of the volume of the bulk soil material. Classes of rock fragment modifiers for round and angular particles larger than 2 mm in diameter are:

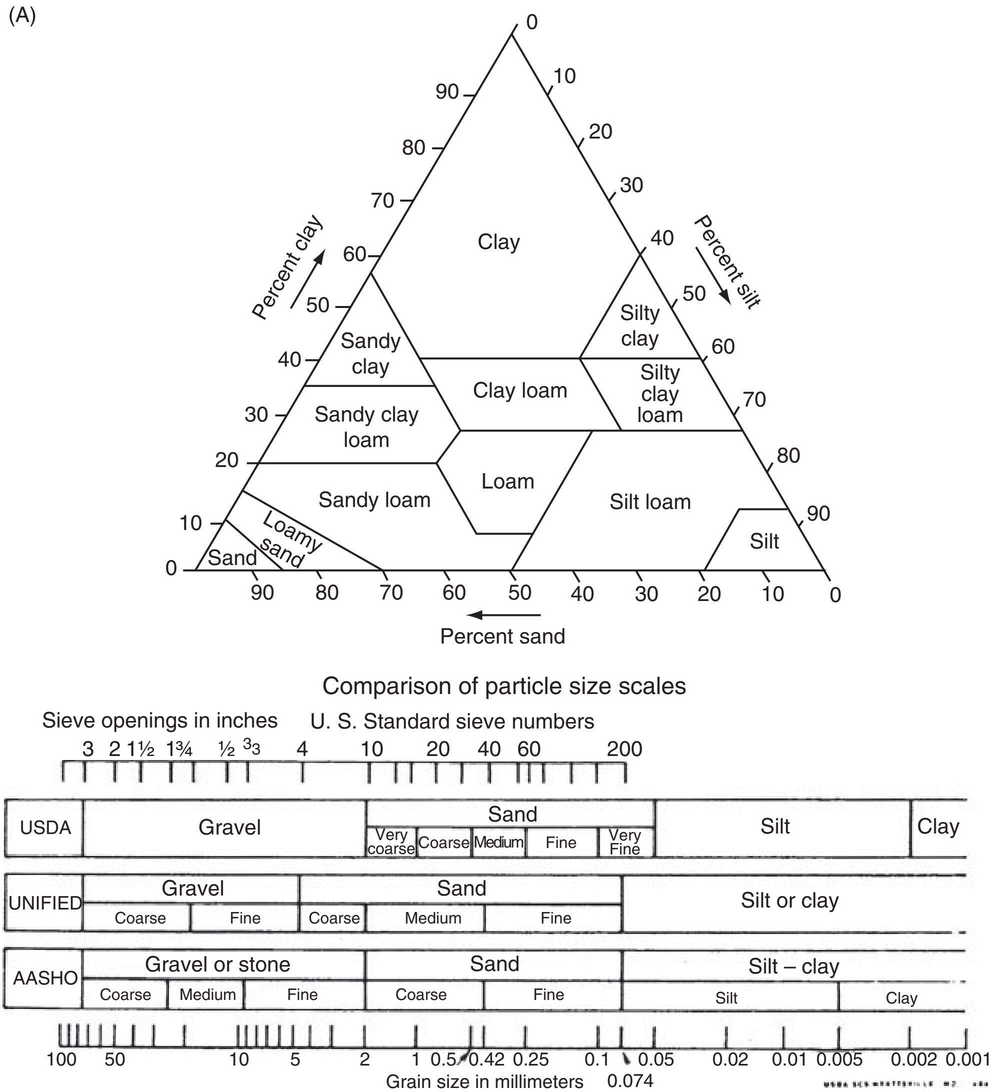

Figure 2.2A. Guide for textural classification of the fine earth fraction (<2 mm).

Size Name	Diameter	Modifier Name
Gravel	2–75 mm	Gravelly
Cobble	75–250 mm	Cobbly
Stone	250–600 mm	Stony
Boulder	>600 mm	Bouldery

For rock fragments that are flat the following names, based on length of their longest axis, are used:

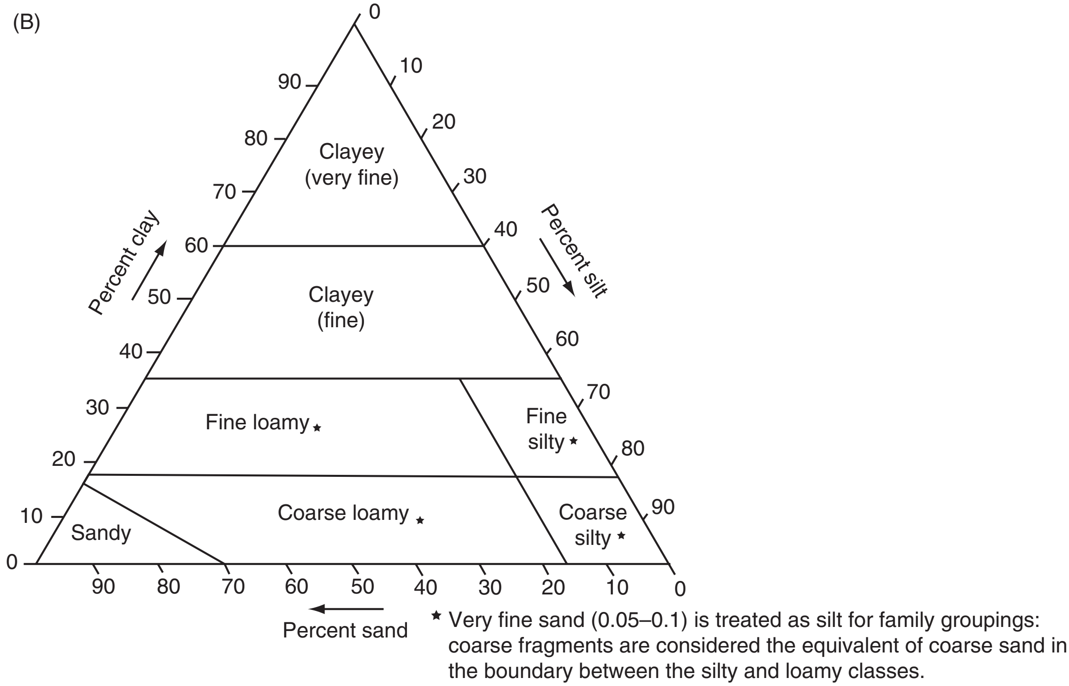

Figure 2.2B. Guide for soil family groupings on the basis of particle size, i.e., particle-size distribution of the whole soil.

Size Name	Longest Dimension	Modifier Name
Channer	2–150 mm	Channery
Flagstone	150–380 mm	Flaggy
Stone	380–600 mm	Stony
Boulder	>600 mm	Bouldery

If the fine earth fraction texture is loam and the soil material contains between 15 and 35% by volume rock fragments and gravel is the dominant size fraction, then the textural class of the material is "gravelly loam." If the soil contains from 35 to 60% gravel, the textural class is "very gravelly loam," and if it contains from 60 to 90% gravel it is classified as "extremely gravelly loam." The same conventions are used for the other rock fragment modifiers of texture (for example, "channery sandy loam," "very flaggy clay loam"). If the soil material contains 90% or more rock fragments, the textural class is named for the dominant rock fragment size class, for example, "cobbles." In all cases, the term describing the quantity of rock fragments is based on the total volume of rock fragments. The rock fragment size modifier is based on the largest dominant fragment size. If there is a mixture of sizes, a smaller size class is named only if the volume percent of the smaller class is more than two times the volume percent of a larger class. For example, a texture with 21% gravel and 10% cobbles is "gravelly," but a texture with 20% gravel and 11% cobbles is "cobbly."

Organic soil materials use terms in place of textural classes based on the degree of organic matter decomposition. *Peat* is organic material that is relatively undecomposed plant tissues; *muck* is highly decomposed organic material, and few plant tissue fibers are observed; *mucky peat* is of intermediate decomposition. To estimate undecomposed tissue or "rubbed fiber" content wet samples are rubbed with the hands and carefully examined to observe undecomposed tissue.

Soil Structure. Structure refers to the aggregation of individual soil particles into larger units with planes of weakness between them. Individual aggregates are known as "peds." There is no technique for observing structure that is applicable to all soils. Carefully probing the exposed soil profile with a knife and prying out volumes of soil while observing how the material crumbles into peds is most often applicable.

Horizons that do not have aggregates with naturally preserved boundaries (peds) are considered to be "structureless." Two forms of a structureless condition are recognized: *single grain* when the soil material is not cohesive and the individual particles are pried or fall from the profile face, or *massive* when the material is cohesive and fails to show any naturally occurring planes of weakness when pried from the profile face.

In horizons where peds are observed, three features of structure are usually described.

Type refers to the shape of peds. *Granular* peds are more or less spherical with little or no accommodation of adjoining peds. *Platy* peds are flat and plate-like, usually oriented horizontally. *Prismatic* peds are elongated in the vertical axis, with flat tops, and horizontally bounded by rather flat ped faces. *Columnar* peds are similar to prismatic peds, but have rounded and commonly bleached (usually due to dispersion by sodium) tops. *Wedge* peds are elliptical lenses that terminate in acute angles and are bounded by slickensides. *Blocky* peds are block-like or polyhedral. Blocky structure with sharp angles between ped faces is referred to as *angular blocky,* and when the angle between faces is rounded, it is termed *subangular blocky*.

Size of the peds is identified in six categories that vary depending on ped type (see Table 2.1) and always refers to the smallest dimension of the ped.

Grade identifies how well formed the peds are and how easily the structure is observed in place in the pedon. Grade of structure is moisture dependent, and considerable experience is often required to evaluate either extremely wet or extremely dry soils. The grade is *weak* if the peds are barely observable in place and when soil material is removed from the horizon and gently bounced up and down in the hands, few identifiable peds survive. The grade is *moderate* if the peds are well-formed and evident in place and most of the material removed from the pedon retains the ped shapes when observed in a hand sample. *Strong* structure is distinct and obvious in place, and peds separate cleanly and persist in a hand sample. These three features of structure are conventionally written in the order of grade, size, and type, for example, weak medium subangular blocky, or moderate thin platy.

Table 2.1. Classes of soil structure

	Shape of Structure		
Size Classes[a]	Granular, Platy[b] (mm)	Prismatic, Columnar, Wedge (mm)	Blocky (mm)
Very fine	<1	<10	<5
Fine	1–2	10–20	5–10
Medium	2–5	20–50	10–20
Coarse	5–10	50–100	20–50
Very coarse	≥10	100–500	≥50
Extremely coarse		≥500	

[a]Size limits refer to the smallest dimension of the structural unit.
[b]Substitute "thin" for "fine" and "thick" for "coarse" for platy size names.

Consistence. Consistence of soil material is an evaluation soil of the cohesive and adhesive properties and of the response of the soil material to applied pressure. Field evaluation of soil consistence includes rupture resistance, stickiness, and plasticity of hand specimens. Soil water content is critical to the assessment of these properties.

Rupture resistance is an evaluation of the strength of soil to withstand an applied stress. To make this evaluation, a block of soil, approximately 2.5–3.0 cm on edge, is removed from the horizon (crusts or plates should be 1.0–1.5 cm long). The moisture content is estimated, and the following operations required to rupture the soil specimen are performed to evaluate the material's rupture resistance class. Note that the test for cementation requires that the specimen be air dried, then submerged in water for at least one hour. Air-dried samples that do not slake (fall apart) when submerged for one hour are cemented.

Class			
Dry	Moist	Cementation	Specimen Fails Under
Loose	Loose	Not applicable	Intact specimen not obtained
Soft	Very friable	Noncemented	Very slight thumb and forefinger pressure
Slightly hard	Friable	Extremely weakly cemented	Slight thumb and forefinger pressure
Moderately hard	Firm	Very weakly cemented	Moderate thumb and forefinger pressure
Hard	Very firm	Weakly cemented	Strong thumb and forefinger pressure
Very hard	Extremely firm	Moderately cemented	Two hands required to rupture block

Extremely hard	Slightly rigid	Strongly cemented	Underfoot with full body weight
Rigid	Rigid	Very strongly cemented	Ruptures when struck with 1 kg hammer
Very rigid	Very rigid	Indurated	Can not be ruptured with hammer

Stickiness is the tendency of wet soil material to adhere to other objects. It is evaluated by wetting a handful of soil with continued squeezing and mixing while alternately adding water and drier soil to determine when the greatest degree of stickiness is achieved. At that point a sample of the material is placed between the thumb and forefinger, squeezed and released. The stickiness class is determined according to the following observations of how the material reacts when the thumb and forefinger are separated:

Stickiness Class	Test Description
Nonsticky	Almost no material adheres to thumb or forefinger
Slightly sticky	Some soil adheres to both fingers; soil stretches little
Moderately sticky	Soil adheres to both fingers; soil stretches some
Very sticky	Soil adheres firmly to both fingers; soil stretches greatly

Plasticity is an estimate of the soil materials ability to deform without rupturing. It is evaluated by molding a handful of soil from which particles greater than 2 mm have been removed, alternately wetting and adding dry soil to obtain maximum plasticity and forming a "roll or wire" 4-cm long that can support its own weight when held on end by rolling the material between the hands. The result is recorded in one of the following plasticity classes:

Plasticity Class	Test Description
Nonplastic	A roll 6-mm thick cannot be formed
Slightly plastic	A roll 6-mm thick supports its own weight
Moderately plastic	A roll 4-mm thick supports its own weight
Very plastic	A roll 2-mm thick supports its own weight

Roots and Pores. Root quantity and size in each horizon, along with observations of the present vegetation at the site, provide some evidence of soil-plant relationships. Roots, especially those of perennial plants, experience the conditions in each horizon over long periods of time and respond to chemical and physical conditions not clearly evident during profile examination.

Pores are the voids in the soil that are filled with soil solution or soil atmosphere, and are the conduits through which soil solution and soil atmosphere move in the soil.

To record root or pore quantity and size classes, the exposed face of each horizon is examined to determine the diameter of the roots or pores exposed and noted according to the following classes. More than one class of root or pore diameters may be present.

Size Class of Roots or Pores	Root or Pore Diameter
Very fine	<1 mm
Fine	1–2 mm
Medium	2–5 mm
Coarse	5–10 mm
Very Coarse	10 mm or larger

After the diameter class or classes have been determined, several "unit areas" in the exposed horizon are examined to determine the quantity of roots and pores present. For the fine and very fine roots or pores the unit area is 1 cm^2. For medium and coarse roots or pores the unit area is 1 dm^2 and for very coarse roots or pores the unit area is 1 m^2. The number of roots or pores observed in the defined unit area is recorded as follows:

Root or Pore Quantity Classes	Number of Roots or Pores Observed
Few	<1 per unit area
Common	1–5 per unit area
Many	5 or more per unit area

For roots, the additional classes of "very few" (<0.2 per unit area) and "Moderately few" (0.2–1 per unit area) may be used. While observing roots it is often useful to note the location of the roots with respect to structural units. For example, it may be noted that roots are confined to ped faces or that roots are uniformly distributed. It is customary to express the presence of roots in a horizon by quantity, size, and location, for example, few coarse roots throughout and many very fine roots on ped surfaces, or many fine roots in a mat at the top of the horizon. Pore shape and continuity can also be described.

Horizon Boundaries. Horizon boundaries are identified as to their distinctness and topography. The distinctness of the horizon boundary is determined by how accurately the investigator feels the depth of the boundary between adjacent horizons can be identified. The following conventions of *distinctness* are used to express the thickness of the transitional zone between horizons:

Distinctness	Transitional zone thickness
Very abrupt	<0.5 cm
Abrupt	0.5–2 cm
Clear	2–5 cm
Gradual	5–15 cm
Diffuse	15 cm or more

Horizon boundaries often do not parallel the surface of the soil. Some horizons are discontinuous. The following conventions are used to express the *topography* of the horizon boundaries observed at the profile exposure:

Horizon topography	Topographic characteristics
Smooth	Planar boundary with few irregularities
Wavy	Boundary undulates with vertical distance less than lateral distance
Irregular	Boundary undulates with lateral distance less than vertical distance
Broken	Horizon is laterally discontinuous

The distinctness and topography are recorded for the lower boundary of each horizon.

Additional Features. Additional features such as ped coatings, concretions, pH, salinity, the presence of carbonates, and any observation the investigator deems of interest and potential value in characterizing the horizon should be noted. The horizon description in the field is the vital link between the real soil and any data that may be obtained from laboratory analysis of samples.

Soil Horizon Designations. After the morphological characteristics of each horizon have been described, it is desirable to identify each horizon with conventional nomenclature for convenient communication. The horizon designation summarizes the investigator's interpretation of the properties observed. Horizon nomenclature is a guide to, but not the same as, diagnostic horizon nomenclature used in *Soil Taxonomy*. Horizon nomenclature has historically undergone several changes. The following section presents abbreviated horizon nomenclature from the *Keys to Soil Taxonomy, 11th ed.* (Soil Survey Staff 2010) that should be consulted for more detailed definitions.

Master Horizons And Layers. Master or major horizons and layers are designated by the following capital letters:

O Horizons or layers dominated by organic soil material: O horizons formed from organic litter derived from plants and animals and deposited on either an organic or a mineral surface. O horizons usually are present at the soil surface, but they may constitute the entire thickness of the soil in the case of organic soils, or may be buried by mineral soil.

L Horizons or layers of organic or mineral limnic material. Limnic materials were deposited in water by aquatic organisms, or were derived from aquatic plants and subsequently modified by aquatic animals. Limnic materials include coprogenous earth, diatomaceous earth, and marl. L horizons and layers are used only in Histosols.

A Mineral horizons at the soil surface or below an O horizon: A horizons have humified organic matter mixed with mineral material resulting from the decomposition of plant residues or have properties resulting from cultivation or pasturing that has physically disturbed the horizon. They also show evidence of obliteration of much or all of the original rock structure, including fine stratification of sediments, and often show evidence of eluviation of clays, salts, carbonates, iron and aluminum compounds, and soluble organic compounds.

E Mineral horizons in which loss of organic matter, clay, iron or aluminum has concentrated sand and silt particles of quartz or other resistant minerals. E horizons have lighter colors than overlying A horizons and underlying B horizons. They may directly underlie an O horizon. E horizons are not present in many soils.

B Mineral horizons formed below O, A, and/or E horizons. The parent material has been significantly altered by (1) accumulation of silicate clay, iron, aluminum, carbonates, gypsum, silica, or humus, (2) removal of carbonates or gypsum, (3) alteration in the form of brittleness, development of soil structure, increased redness of soil color, gleying, or residual concentration of oxide compounds. There are many kinds of B horizons, but the main consideration in identifying a B horizon is that it formed as subsoil, below one or more horizons and is significantly different from the material in which if was formed as a result of pedogenic processes.

C Mineral horizons or layers, other than strongly cemented and harder bedrock, with little or no alteration by soil forming processes. C horizons or layers lack properties of O, A, E, or B horizons. Plant roots such as tap roots of trees or a few fine roots that leave only trace amounts of organic carbon upon their death may be present at widely spaced intervals in some C horizons.

R Layers of bedrock that are strongly cemented to indurated. Rock material that is of sufficient hardness that hand digging with a spade is impractical even when the material is moist is designated R. Small cracks, partially or totally filled with soil material and occupied by roots, are frequently present in R layers.

M Root-limiting subsoil layers consisting of nearly continuous, horizontally oriented, human-manufactured materials. Asphalt, concrete, and geotextile liners are examples.

W Water layers within or beneath the soils. The water may be liquid (for example, water below organic soil material in a floating bog) or permanently frozen (for example, an ice lens in a soil with permafrost). It is not used to designate shallow water, snow, or ice above the soil surface.

Transitional and Combination Horizons. Where a horizon of substantial thickness is present between two master horizons, a transitional or combination horizon may be described. Transitional horizons, which are dominated by properties of one master horizon while having subordinate properties of an adjacent master horizon are designated by two master horizon capital letters. The first letter indicates the dominant master horizon characteristics, and the second letter indicates the subordinate horizon characteristics. For example, an AB horizon indicates a transitional horizon between the A and B horizon that is more like the A horizon than the B horizon. A BA horizon is a transitional horizon between the A and B master horizons that is more like the B horizon than the A horizon. Other commonly designated transitional horizons include AE, EA, EB, BE, BC, CB, and AC.

Combination horizons are recognized where distinct parts of two master horizons are recognizable in the horizon and one of the components surrounds the other. Such combination horizons are designated as A/B, B/A, E/B, B/E, or B/C. The first symbol designates the material of greatest volume in the horizon.

Subordinate Distinctions Within Master Horizons and Layers. Lowercase letters are used to designate specific features within master horizons and layers. O, L, and B horizons must have at least one subordinate distinction. Many of the designations include the term "accumulation," which means that the horizon contains more of the material in question than is presumed to have been present in the parent material.

a Highly decomposed organic material: Used with the **O** master horizon with rubbed fiber content less than 17% by volume.

b Buried genetic horizon: This designation is used only if the buried mineral horizon contains clearly identifiable features of a genetic soil horizon. It is not used in organic soils or to identify a buried **O** horizon.

c Concretions or nodules: Indicates a significant accumulation of iron, aluminum, manganese, or titanium cemented concretions or nodules. The hardness or consistence of the concretion or nodule should be described.

co Coprogenous earth: A limnic layer of coprogenous earth. Used only with L.

d Physical root restriction: This symbol is used to indicate noncemented, root-restricting, naturally occurring, or human-made layers such as basal till, plow pans, and other mechanically compacted zones. Roots do not enter except along fracture planes.

di Diatomaceous earth: A limnic layer of diatomaceous earth. Used only with L.

e Organic material of intermediate decomposition: Used with **O** horizons with rubbed fiber content of 17 to 40% by volume.

f Frozen soil or water: This is used for horizons or layers that contain permanent ice.

ff Dry permafrost: Indicates a horizon or layer that is continually colder than 0°C and is not cemented by ice.

g Strong gleying: This symbol is used with B and C horizons to indicate either that iron has been reduced and removed (redox depletion) or that iron has been reduced under stagnant saturated conditions and is preserved in the soil (reduced matrix). Strongly gleyed horizons and layers have dominant colors with chroma of 2 or less, or have gley hues (for example, GY, G, BG, or B); many of these horizons and layers also have redox concentrations. The **g** is not used in materials with low chroma that have no evidence of saturation or history of wetness, as in many E horizons, or in C horizons developed from parent materials with low chroma.

h Illuvial accumulation of organic matter: Used only in B horizons, the h indicates an illuvial accumulation of amorphous, dispersible organic matter-sesquioxide complexes. The sesquioxides are dominated by aluminum, but are present in very small quantities.

i Slightly decomposed organic material: Used with O horizons with rubbed fiber content of 40% or more by volume.

j Accumulation of jarosite: Used mostly with B and C horizons to indicate the presence of jarosite, $KFe_3(SO_4)_2(OH)_6$, resulting from the oxidation of sulfur.

jj Evidence of cryoturbation: Used mostly with B and C horizons to indicate churning of soil materials by freezing and thawing.

k Accumulation of secondary carbonates: Used with B and C horizons, to indicate an accumulation of pedogenic carbonates, usually calcium carbonate (less than 50% by volume).

kk Engulfment of horizon by secondary carbonates: Used with B horizons when the horizon fabric is plugged with pedogenic calcium carbonate (50% or more by volume).

m Cementation or induration: Used mostly with B and C horizons to indicate that more than 90% of the horizon is cemented and roots penetrate only through cracks. The cementing material is identified by the appropriate letter, such as kkm or km = carbonate, qm = silica, sm = iron, yym = gypsum, kqm = both carbonate and silica, zm = salts more soluble than gypsum.

ma Marl: Indicates a limnic layer of marl. Used only with L.

n Accumulation of exchangeable sodium: This symbol is used on any master horizon to indicate an accumulation of exchangeable sodium.

o Residual accumulation of sesquioxides: This symbol is used mostly with a B horizon having properties indicative of residual sesquioxides (for example, gibbsite, goethite, and hematite) after intense weathering.

p Tillage or other cultivation disturbance: Indicates disturbance by plowing or other tillage activities or by intensive pasturing. The symbol p is used with O to indicate plowed organic soil material and with A for mineral horizons, even if the materials mixed by the cultivation are from an E, B, or C horizon.

q Accumulation of silica: This symbol is used with B and C horizons where secondary silica (opal) has accumulated.

r Weathered or soft bedrock: This symbol is only used with the **C** layer to indicate layers of bedrock that are moderately cemented to extremely weakly cemented. It designates material including saprolite and partially consolidated sedimentary rock that is hard enough that roots only penetrate along cracks, but soft enough that it can be dug with a spade or shovel.

s Illuvial accumulation of sesquioxides and organic matter: This symbol is only used with B horizons to indicate the accumulation of illuvial amorphous, dispersible complexes of sesquioxides and organic matter if both the organic matter and the sesquioxides components are significant and if the color value or chroma, moist, is 4 or more. It is used in conjunction with h (Bhs) if the color value and chroma, moist, are 3 or less.

ss Presence of slickensides: Slickensides are formed by shear failure as clay minerals swell upon wetting. Slickensides are an indicator of vertic characteristics.

t Accumulation of silicate clay: Indicates accumulation of clay, either formed within the horizon and subsequently moved within the horizon, or moved by illuviation into the horizon, or both. Evidence of clay illuviation or migration within the horizon includes clay coatings on ped surfaces or in pores, as lamellae, or as bridges between mineral grains. Although usually used in B horizons, it may be used in C and R layers.

u Presence of human-manufactured materials (artifacts): Indicates the presence of manufactured artifacts such as processed wood products, bricks, glass, asphalt, and plastic.

v Plinthite: Indicates the presence of iron-rich, humus-poor, reddish material that is firm or very firm when moist and is less than strongly cemented. It usually occurs in a reticulate mottling pattern of red, yellow, and gray colors. Upon exposure, as in a road cut, the material hardens irreversibly with repeated wetting and drying. This symbol is used with B and C.

w Development of color and structure: This symbol is used with B horizons that have developed color, usually redder than the A or C materials, or soil structure, but do not have apparent illuvial accumulations.

x Fragipan character: This symbol is used with B and C to designate genetically developed firmness and brittleness, and often a higher bulk density than adjacent layers, but no cementation. Some part of the layer is root-restrictive.

y Accumulation of gypsum: This symbol is used with B and C to indicate pedogenic gypsum accumulations when the horizon fabric is dominated by minerals other than gypsum.

yy Dominance of horizon by gypsum: Indicates that the horizon fabric is dominated by the presence of gypsum (usually 50% or more by volume), and pedogenic and/or lithologic features are disrupted by gypsum crystal growth.

z Accumulation of salts more soluble than gypsum: This symbol is used with B and C to identify accumulation of salts more soluble than gypsum (for example, halite, thenardite, and trona).

Some horizons have features of more than one of the subordinate distinctions. If more than one suffix is used, the following letters are written first: a, d, e, h, i, r, s, t (except as Crt), and w. Except in Bhs or Crt horizon, none of these letters is used in combination in a single horizon. If more than one suffix is needed and the horizon is not buried, the following symbols are written last: c, f, g, m, v, and x. If a horizon is buried, the suffix b is written last, and is used only for buried mineral soils.

In practice it is frequently desirable to further subdivide master, transitional, and subordinate horizons. This situation commonly occurs when it is desirable to sample by small increments of depth. Arabic numbers are added as suffixes to the horizon designations to identify vertical subdivisions within a horizon when a horizon or layer is identified by a single letter or by the same combination of master and subordinate distinctions. For example, an A1-A2-AB-Bt1-Bt2-Btk-BC-C horizon sequence indicates two A subhorizons and two Bt subhorizons. Note that the Btk horizon does not include a numerical suffix.

Discontinuities. Layers or horizons in a profile that are interpreted by the investigator to result from geologic processes are referred to as discontinuities and identified by a number before the master horizon identifier. Most discontinuities are observed as distinct changes in particle-size distribution. A discontinuity is noted to reflect the investigator's interpretation of the genesis of the layer in question. Some discontinuities are common to many soils within a region such as loess over glacial drift. Loess is gravel free, except for local mixing, and most glacial drift has some gravel. A sequence of horizons such as Ap-E-Bt1-2Bt2-2BC-2C indicates that the top three horizons are from one geologic material, and a material of different origin is present in the lower three horizons described. Note that the suffix number identifying subdivisions of the Bt continue through the discontinuity, and a 1 is not used to identify the top geologic layer.

By convention organic layers are not designated as lithologic discontinuities. Also, buried layers that have properties indicative of pedogenic development are designated as buried horizons (i.e., b subordinate distinction), and may or may not be designated as discontinuities. If the buried horizon is developed in material lithologically similar to the overlying material, a discontinuity usually is not identified. However, if the materials are lithologically different, a discontinuity is indicated.

Volcanic ash presents problems in identifying discontinuities. Ash from successive eruptions often blankets existing topography, thus discontinuities conform to present topography. Also, any given eruption event may deposit rather contrasting materials. Many alluvial deposits are also highly stratified. In these cases, lithologic discontinuities generally are indicated only if particle-size distribution differs markedly from layer to layer and gives rise to strongly contrasting particle-size classes.

Use Of The Prime and Caret Symbols. Occasionally two horizons may develop in a profile and have the same combination of letters and numbers for horizon designations, but are separated by an unlike horizon. Where the investigator interprets this to be due to pedogenic processes, the lower of the two horizons is designated as prime ('). For example, an A-E-Bh-E'-Bt profile indicates an E horizon formed in association with a Bt horizon and subsequently a Bh horizon has formed within the E horizon. Soils with such genetic scenarios are referred to as bisequal.

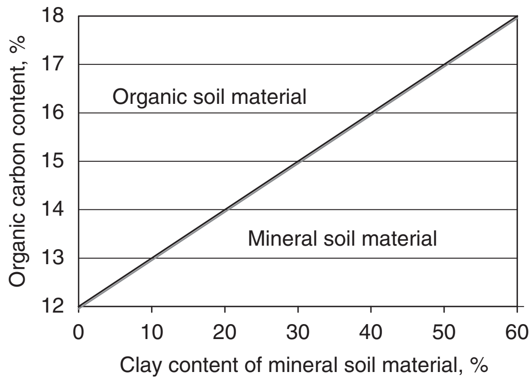

Figure 2.3. Organic carbon and clay contents required to differentiate organic and mineral soil materials.

The caret symbol (ˆ) is used as a prefix to master horizon designations to indicate the presence of human-transported materials, for example, fill material at a construction site or garbage disposal land-fill site. The horizon sequence ˆA-ˆC-Ab-Btb1-Btb2-C indicates that the upper two horizons are formed in materials that were transported by humans and that buried a soil with an A-B-C horizon sequence.

Diagnostic Soil Materials and Horizons For Classification. Quantified identification of different soil materials and horizons is an integral part of *Soil Taxonomy*. Layers or horizons in a soil that have specifically defined properties are known as diagnostic soil materials or horizons. Most diagnostic soil material and horizon definitions require some laboratory analyses for identification and are not intended to conform to soil horizon designations presented in the previous section. However, it is prudent for the soil scientist to be aware of the criteria used to identify diagnostic materials and horizons when describing and sampling a soil profile. A discussion of the rationale used in the development of diagnostic soil materials and horizons is available (Smith 1986).

Definition of Soil Materials. All soil materials are identified as either mineral or organic soil materials. Both field description and laboratory analysis are required to identify many of the soil materials.

Mineral soil materials are either saturated with water for less than 30 cumulative days per year in normal years and contain less than 20% organic carbon by weight, or are saturated with water 30 or more cumulative days per year in normal years, unless artificially drained and after excluding live roots have an organic carbon content, by weight, of one of the following:

a. Less than 18% C if the mineral fraction contains 60% or more clay; or
b. Less than 12% C if the mineral fraction contains no clay; or
c. Less than 12 + (clay percentage multiplied by 0.1) percent C if the mineral fraction contains less than 60% clay (Figure 2.3).

Organic soil materials contain more organic carbon than specified for mineral soil material (Figure 2.3).

Named Mineral Soil Materials. Most of the following are mineral soil materials but some, as indicated, also apply to organic soil materials. These terms are used to identify various soils in *Soil Taxonomy*.

Albic materials — Soil materials with a white to gray color mainly due to the color of primary sand and silt particles: (1) Chroma is 2 or less, moist value is 3 and dry value is 6 or more *or* moist value is 4 or more with a dry value of 5 or more, *or* (2) Chroma is 3 or less, moist value is 6 or more or dry value is 7 or more, *or* (3) Chroma is controlled by the color of uncoated silt or sand grains, the hue is 5YR or redder, and the color values are as listed in item 1 above. These colors imply that organic matter, clay, and/or iron oxides have been removed from the material. Light-colored volcanic or sedimentary deposits are not considered albic materials.

Andic materials (Properties) — Andic soil properties are mainly due to the presence of significant amounts of short-range-order compounds, including allophane, imogolite, and ferrihydrite. Most andic materials are formed from volcanic ash or other volcanic materials that contain volcanic glass. They have less than 25% organic carbon by weight. If phosphate retention is 85% or more, the ammonium oxalate extractable aluminum plus 1/2 of the iron percentage (Alo + 1/2 Feo) totals 2.0 or more and bulk density at $-33\,\text{kPa}$ water potential is $0.90\,\text{g cm}^{-3}$ or less. If phosphate retention is 25% or more and 30% or more of the particles are 0.02 to 2.0 mm in diameter one of the following conditions is present:

1. (Alo + 1/2 Feo) percentage totaling 0.40 or more and in the 0.02 to 2.0 mm fraction, 30% or more volcanic glass, or
2. (Alo + 1/2 Feo) percentage totaling 2.0 or more and in the 0.02 to 2.0 mm fraction, 5% or more volcanic glass, or
3. (Alo + 1/2 Feo) percentage totaling between 0.40 and 2.0 and in the 0.02 to 2.0 mm fraction, a volcanic glass percentage intermediate between 5 and 30% so that ([Alo + 1/2 Feo] percentage times 15.625) plus (volcanic glass percentage) = 36.25 or more (Figure 2.4).

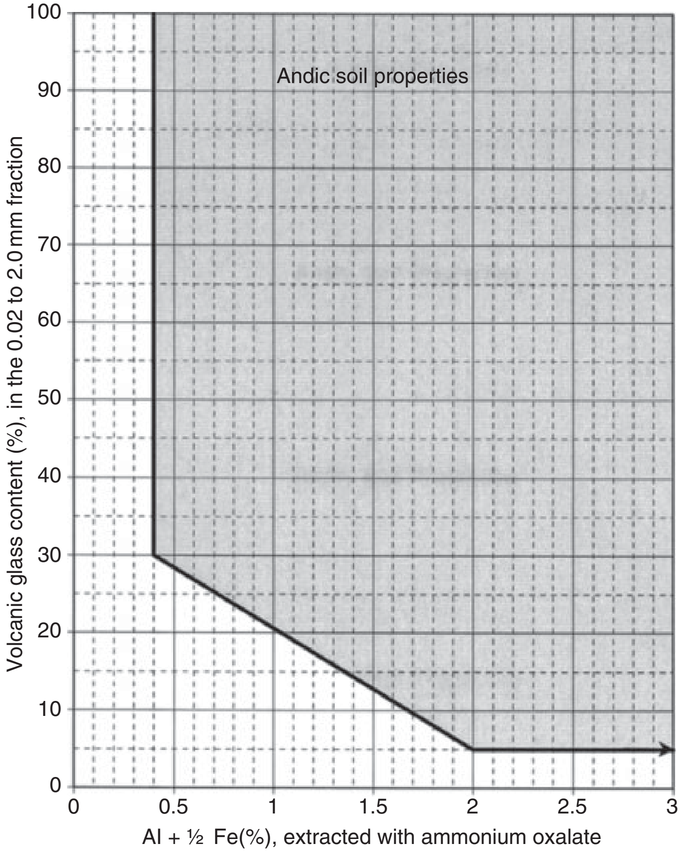

Figure 2.4. Definition of andic soil properties (shaded area) based on glass content and quantity of acid-oxalate-extractable aluminum and iron (Soil Survey Staff 2010).

Densic materials	Root restricting material that is relatively unaltered by pedogenic processes but not cemented. Roots enter along cracks at horizontal spacing of 10 cm or more.
Durinodes	These are weakly cemented to indurated nodules or concretions with a diameter of 1 cm or more that are cemented by SiO_2 (opal). When peds are air dry they do not slake in water or HCl but are destroyed by hot KOH after acid washing.
Gelic materials	Mineral or organic materials that show evidence of churning by freezing and thawing (cryoturbation) in the seasonally thawed layer (active zone) above the permafrost table.
Identifiable Secondary Carbonates	Authigenic carbonate that has been translocated within the soil and is normally present as concretions, soft masses, or filaments, or as coatings on ped surfaces, the underside of rock fragments, or walls of large pores.

Paralithic materials	Partially weathered bedrock or weakly consolidated bedrock such as sandstone, siltstone, or shale and other isovolumetrically weathered bedrock that retains the fabric of the consolidated rock. The materials are extremely weakly cemented to moderately cemented. Roots enter only in cracks at intervals greater than 10 cm.
Permafrost	A thermal condition of material that remains below 0°C for two or more years in succession.
Spodic materials	Mineral materials dominated by illuvial amorphous materials of organic matter and aluminum with or without iron. Spodic materials have a pH value in 1:1 water of 5.9 or less and an organic carbon content of 0.6% or more. Spodic materials often underlie albic material in the pedon and are identifiable by darker and redder colors. Normally they have an optical density of oxalate extract (ODOE) value of 0.25 or more, and that value is at least two times as high as the ODOE value of an overlying eluvial horizon, or they have (Alo + 1/2 Feo) percentage totaling 0.50 or more and that value is at least two times as high as is an overlying eluvial horizon. The ODOE is measured with a spectrophotometer at 430 nm.
Sulfidic materials	Mineral or organic materials that contain oxidizable sulfur compounds with a pH value in 1:1 water more than 3.5, which acidifies by 0.5 or more pH units to a value of 4.0 or less when kept under moist aerobic conditions at room temperature for 16 weeks.

Named Organic Soil Materials. The following terms apply only to organic soil materials.

Fibers	Pieces of plant tissue, excluding live roots, smaller than 2 cm after crushed and shredded in the hands and large enough to be retained on a 100-mesh sieve (openings 0.15 mm). Coarser pieces of wood that cannot be hand crushed are considered coarse fragments. Crushing and rubbing wet organic soil material in the hands is considered the best estimate of the degree of organic decomposition. A small volume of wet material is rubbed between the thumb and forefinger about 10 times before estimating the fiber content of an organic soil material.
Fibric soil materials	Slightly decomposed organic soil material with three-fourths or more of the volume composed of fibers after rubbing, or two-fifths or more by volume rubbed fibers and soil color value and chroma of 7/1, 7/2, 8/1, 8/2/ or 8/3 of a sodium pyrophosphate extract on white filter paper.

Hemic soil materials — Partially decomposed organic soil material that generally has rubbed fiber content between one-sixth and two-fifths by volume, and has sodium pyrophosphate extract colors that do not meet criteria for fibric or sapric materials.

Humilluvic materials — An accumulation of colloidal organic material or humus, 2 cm or more thick, at depth in an organic soil that has been drained and cultivated. Humilluvic materials are usually present immediately above a sandy mineral layer.

Limnic materials — Materials deposited under water by the action of aquatic organisms such as algae or diatoms. Although some of the materials do not meet the organic soil material definition, they are often present in soils dominated by organic soil materials. Three types of limnic materials are recognized:

1. *Coprogenous earth*: Layer of finely divided organic material with a moist color value of 4 or less, containing many fecal pellets, and often present in layers indicating annual deposition in seasonally frozen lakes.
2. *Diatomaceous earth*: Contains organic coated diatoms identifiable upon examination of a dried sample with a 440X microscope. Most diatomaceous earth is mineral soil material.
3. *Marl*: Carbonate-rich limnic material with a moist color value of 5 or more that reacts with dilute HCl. Most marl materials are mineral soil material.

Sapric soil materials — The most decomposed organic soil material that has rubbed fiber content less than one-sixth by volume and specific sodium pyrophosphate extract colors indicating more advanced stages of organic matter decomposition.

Diagnostic Horizons. The following brief definitions are abstracted from the *Keys to Soil Taxonomy, 11th ed.* (Soil Survey Staff 2010) that should be consulted for complete definitions.

Epipedons. Epipedons are simply the uppermost soil horizons. “Epipedon” is not synonymous with A horizon and may be thinner than the A horizons or include some portion of the E, B and/or C horizons designated in field descriptions.

*Anthropic Epipedon*s — An anthropic epipedon is a surface horizon like the mollic epipedon but contains 1,500 or more milligrams of phosphate (P_2O_5), soluble in 1% citric acid, per kilogram. Anthropic epipedons are usually present in small areas were phosphate has been accumulated from human and/or animal waste disposal.

Folistic epipedon Organic material on the soil surface that is saturated less than 30 days during normal years. The folistic epipedon is 20 cm or more thick if it contains more than 75% by volume *Sphagnum* fibers or has a bulk density less than 0.1 g cm^{-3}; otherwise the folistic epipedon is at least 15 cm thick. If plowed (an Ap horizon) or mixed to 25 cm it contains 16% or more organic carbon if it contains 60% or more clay, or 8% or more organic carbon if it contains no clay, or 8 + (clay percentage divided by 7.5) percent or more organic carbon if clay content is between 0 and 60%.

Histic Epipedon A histic epipedon is a surface horizon that is saturated for at least 30 days in most years unless artificially drained. Most are 20- to 40-cm thick, but can be 20- to 60-cm thick if they contain more than 75% by volume *Sphagnum* fibers or if the bulk density moist is less than 0.1 g cm^{-3}. If mixed to a depth of 25 cm histic epipedons must contain 16% or more organic carbon if the mineral fraction contains 60% or more clay or have an organic carbon content of (8 + [percent clay/7.5]) percent or more.

Melanic Epipedon The melanic epipedon is a 30 cm or thicker, black horizon at or near the surface that has andic (see diagnostic mineral soil properties) soil properties, a moist color value and chroma of 2 or less, averages 6% or more organic carbon with 4% or more carbon in all layers. It has a 1.70 or less melanic index. (Add 25 ml of 0.5% NaOH solution to 0.5 g of air dry soil; shake for 60 minutes; add one drop 0.1% Superfloc; centrifuge; pipette 1 ml to test tube; add 20 ml of 0.1% NaOH and mix. Absorbance at 450 nm divided by absorbance at 520 nm equals melanic index.)

Mollic Epipedon A mollic epipedon is a dark colored, organic carbon rich (generally at least 0.6% organic carbon), mineral surface horizon that when mixed to a depth of 18 cm has color values of 5 or less dry, and 3 or less moist, and chroma of 3 or less moist. Structural units have a diameter less than 30 cm and a moderately hard or softer rupture resistance class. Base saturation is 50% or more as measured by ammonium acetate at pH 7 and some part of the epipedon is naturally moist 3 or more months each year when soil temperatures at 50 cm depth exceed 5°C. Soluble phosphate (P_2O_5) content in 1% citric acid is less than 1,500 milligrams (660 milligrams P) per kilogram.

Ochric Epipedon An ochric epipedon is a surface horizon that fails to meet criteria for any of the other epipedons.

Plaggen Epipedon A plaggen epipedon is a surface horizon on a locally raised land surface that is 50 cm or more thick, contains at least 0.6% organic carbon, has color value of 4 or less moist, 5 or less dry, and color value of 2 or less. The plaggen epipedons is created by years of manure and transported soil additions. Plaggen epipedons contain fragments of pottery or other human artifacts, or spade marks below a depth of 30 cm. Plaggen epipedons are very limited in extent occurring most often around medieval settlements in Europe.

Umbric Epipedon An umbric epipedon is a surface horizon like the mollic epipedon but is less than 50% base saturated as measured by ammonium acetate at pH 7.

Subsurface Diagnostic Horizons. Subsurface diagnostic horizons form below an epipedon but may be exposed at the surface by truncation of the pedon. Most are B horizons, but some include A, E, and C horizons as described in the field.

Agric horizon Agric horizons form directly under the plow layer, and are 10 cm or more thick. Silt, clay, and humus accumulations in 5% or more of the horizon that have moist value of 4 or less and chroma of 2 or less, occurring as coatings at least 2-mm thick in earthworm channels or as lamellae at least 5-mm thick.

Albic horizon A light-colored eluvial horizon, 1-cm or more thick, of albic materials (see albic materials).

Argillic horizon An illuvial horizon that contains 1.2 times as much clay as some horizon above or 3% (absolute) more clay content if the horizon above has less than 15% clay or 8% more clay if the horizon above has more than 40% clay. It contains clay that has moved from another horizon or within the horizon. Clay films (argillans) should be detected as pore linings, on ped faces, or as bridges between sand grains in some part of the horizon. It is at least 1/10 as thick as all overlying horizons or more than 15 cm, whichever is thinner, but not less than 7.5-cm thick. Argillic horizons 30-cm or more thick have apparent cation exchange capacity (CEC) values greater than those of by the kandic horizon.

Calcic horizon A noncemented horizon of secondary carbonate accumulation, usually of calcium carbonate (sometimes Mg), generally with at least 15% calcium carbonate equivalent, at least 15-cm thick, and at least 5% more carbonate than an underlying layer.

Cambic horizon	A subsoil horizon of very fine sand, loamy very fine sand or finer texture, 15-cm or more thick with some weak indication of alteration in the form of color development (either more red than the parent material or gray due to iron reduction), development of soil structure, removal of carbonates or gypsum, or accumulation of clay or humus-metal complexes that is insufficient for either an argillic or spodic horizon.
Duripan	A subsurface horizon that is cemented in more than 50% of the volume by opaline silica. Air-dry fragments do not slake in water or HCl, but do slake in hot concentrated KOH.
Fragipan	Subsoil layers 15 cm or more thick that show evidence of pedogenesis, have high bulk density, are brittle when moist, and firm or firmer when moist. The layer is not effervescent in HCl. Air-dry fragments slake when immersed in water.
Glossic horizon	Glossic horizons are 5-cm or more thick and usually occur between an overlying albic horizon and an underlying argillic, kandic, or natric horizon or fragipan. Albic materials constitute 15 to 85% of the horizon, and material like that of the underlying horizon constitute the remainder.
Gypsic horizon	Horizons of gypsum accumulation 15 cm or more thick, containing 5% or more gypsum and 1% visible gypsum. The product of horizon thickness in cm and gypsum percentage by weight is 150 or greater.
Kandic horizon	A subsurface horizon with a texture finer than that of an overlying horizon. It is 30-cm or more thick, composed mainly of low activity clays, so that when the CEC of the soil by ammonium acetate at pH 7 is divided by the percent of clay present the apparent CEC_7 of the clay is 16 cmol kg^{-1} clay or less *and* when the neutral salt CEC of the soil (ammonium acetate extractable bases plus KCl extractable Al) is divided by the percent clay, the apparent effective cation exchange capacity (ECEC) is 12 cmol kg^{-1} clay or less. The increase in clay content at the upper boundary is more than 4% (absolute) if clay content of the overlying 18 cm layer is less than 20%, or 1.2 times the overlying clay content if the overlying clay content is 20 to 40%, or more than 8% (absolute) if the overlying clay content is greater than 40%. The clay content increase at the upper boundary is within a vertical distance of less than 15 cm (i.e., an abrupt, clear, or gradual horizon boundary).

Natric horizon Meets the requirements of an argillic horizon and also has prismatic or columnar structure, or blocky structure with tongues of eluvial material that penetrate the upper 2.5 cm of the horizon. Generally, the exchangeable sodium percentage (ESP) measured at pH 7 is 15 or more, or the sodium adsorption ratio (SAR) is 13 or more, within 40 cm of the upper boundary of the horizon.

Ortstein A layer of spodic materials (see spodic horizon) more than 25-mm thick, in which more than 50% is cemented.

Oxic horizon A highly weathered horizon 30-cm or more thick, with a sandy loam or finer texture, and less than 10% weatherable minerals in the 50- to-200-micron fraction. It has a high content of low activity 1:1 clays, with an apparent ECEC value of 12 or less cmol kg^{-1} clay *and* 16 or less cmol kg^{-1} clay by ammonium acetate at pH 7 (similar to a kandic horizon). Either the clay content does not increase with increasing depth or the increase in clay content occurs over a diffuse horizon boundary, in contrast with the clay increase requirements for a kandic horizon.

Petrocalcic horizon Carbonate-cemented horizons 10 cm or more thick, or at least 1-cm thick if it directly overlies bedrock. If cracks are present, the horizontal spacing is 10 cm or more.

Petrogypsic horizon Horizons 5-mm or more thick, cemented with secondary gypsum, with at least 40% gypsum by weight. Roots penetrate only along cracks spaced at horizontal intervals of 10 cm or more.

Placic horizon A thin, black to dark reddish layer cemented by iron and/or manganese and organic matter. The placic horizon is at least 1 mm thick, and, if associated with spodic material, is less than 25-mm thick. (See Ortstein.) Roots can penetrate along vertical cracks spaced at least 10-cm apart.

Salic horizon A horizon of accumulation of salts more soluble than gypsum, at least 15-cm thick. Soluble salts in soil are subject to water movement, therefore, the salic horizon must have an electrical conductivity (EC) of at least 30 dS m^{-1} for 90 consecutive days or more in normal years. The product of horizon thickness in cm and the EC in dS m^{-1} is 900 or more.

Sombric horizon A subsoil horizon of illuvial humus, but without associated Al as in the case of a spodic horizon. The base saturation is less than 50% by ammonium acetate. The sombric horizon has a lower color value or chroma, or both, than the overlying horizon and usually contains more organic carbon. These horizons have been identified in well-drained soils in cool, moist soils in tropical areas.

Spodic horizon	An illuvial layer containing at least 85% spodic material (see spodic materials), is at least 2.5-cm thick, and is not part of an Ap horizon.
Sulfuric horizon	Mineral or organic soil horizon 15 cm or more thick that has a pH value of 3.5 or less due to the presence of sulfuric acid as evidenced by the presence of yellow jarosite concretions or more than 0.05% water-soluble sulfate.

Other Diagnostic Soil Characteristics. In addition to diagnostic materials and horizons, several other features are named and used as diagnostic criteria in *Soil Taxonomy*.

Abrupt textural change	Refers to a boundary between an ochric epipedon or albic horizon and an underlying argillic horizon. If the ochric epipedon or albic horizon contains less than 20% clay, clay content doubles in a vertical distance of 7.5 cm or less. If the ochric epipedon or albic horizon has 20% or more clay, an absolute clay content increase of at least 20% clay is required within a vertical distance of 7.5 cm or less, and the clay content in some part of the argillic horizon is at least two times the clay content in the overlying horizon.
Anhydrous Conditions	Subsoil condition where the temperature remains below 0°C, but due to a lack of water, it is not cemented by ice. Also known as dry permafrost.
COLE (coefficient of linear extensibility)	A measure of the expansion and shrinkage as a soil material wets and dries. COLE values are expressed as a ratio of the difference between moist (33 kPa) length (Lm) and dry length (Ld) of a clod divided by the dry length of a clod, that is, ([Lm – Ld)/Ld], measured in the same plane of the clod.
Fragic Soil Properties	Material with brittle properties of a fragipan, but lacking the thickness requirements of a fragipan.
Free Carbonates	Soil carbonates that effervesce with HCl. There is no implication of pedogenesis.
Glacic Layer	Massive ice or ground ice in lenses or wedges. The layer is 30-cm or more thick and contains 75% or more visible ice.
Interfingering of Albic Material	Term used to identify 5 cm or greater vertical intrusions of albic horizon material into underlying argillic, kandic, or natric horizons. Intrusions are between adjacent peds, at least 2-mm wide and constitute less than 15% of the volume of the horizon they penetrate.

Lamellae — A number of thin clay-enriched layers separated by coarser textured, usually sandy layers, in the subsoil. The clay in each lamella is present as bridges between sand grains. A lamella is less than 7.5-cm thick and nearly horizontal but may undulate and not be continuous throughout a pedon. If there are enough lamellae present, an argillic, kandic, or natric horizon is identified. Lesser amounts of lamellae are indicative of certain cambic horizons.

Linear Extensibility — The product of the COLE value and the thickness in cm of the layer being measured, that is (COLE value × layer thickness, [cm]) equals linear extensibility of that layer.

Lithologic Discontinuities — A layer in a soil profile that results from a geologic process rather than a pedogenic process is known as a lithologic discontinuity. No quantitative criteria have been found to precisely identify all lithologic discontinuities. Several features that indicate a lithologic discontinuity are abrupt textural contact, contrasting sand sizes, stone lines, color or composition of minerals, shape and orientation of rocks, and detailed particle-size evaluation on a calculated clay-free basis. The best line of evidence to determine lithologic discontinuities comes from examination of the layers in question throughout the landscape. If the feature is related to present landscape relief it is probably pedogenic, except for volcanic ash deposits. If it is not related to present relief it is more likely to be geologic.

n *Value* — *n* value is a prediction of a saturated soils ability to support grazing livestock or other trafficking loads. $n = (A - 0.2R)/(L + 3H)$ where A is the percentage of water in the field condition; R is the percentage of silt plus sand; L is the percentage at clay; and H is the percentage of organic matter. Values greater than 0.7 indicate weak bearing strength and are approximated by squeezing a handful of wet soil in the field. If the material oozes between the fingers weak bearing strength is indicated. High *n* values indicate severe subsidence if artificially drained.

Plinthite — Plinthite is humus-poor and iron-rich soil material that hardens irreversibly if exposed to repeated wetting and drying *in place*. Aggregates removed from the pedon and repeatedly wetted and dried seldom harden irreversibly. Plinthite forms by alternating oxidation and reduction reactions in a soil. This most readily occurs as a water table fluctuates through a layer of soil material containing some iron oxide. A mottled pattern of red, iron-enriched bodies, and gray to white bodies of iron depletion develops. The volume of the iron-enriched material present in a soil horizon is a criterion for classifying the soil.

Resistant Minerals Resistant minerals are silt- and sand sized minerals that withstand decomposition in humid climates. Quartz is the most common of these minerals; other examples include rutile, zircon, tourmaline, anatase, iron oxides and oxyhydroxides, gibbsite, and kandites.

Volcanic Glass Optically isotropic transparent volcanic materials of any color including glass, pumice, glass-coated crystalline minerals, glass aggregates, and glassy materials.

Weatherable Minerals Weatherable minerals are relatively unstable soil minerals in humid climates. They include sepiolite, talc, glauconite, and all 2:1 phyllosilicate minerals except hydroxy interlayered minerals (HIM). All sand- and silt-sized feldspars, feldspathoids, ferromagnesian minerals, volcanic glass, dolomite, zeolites, and apatite are considered weatherable minerals.

Diagnostic Contacts. A "contact" is a depth in the soil profile where roots are physically restricted from entering except at infrequent lateral intervals of more than 10 cm. This density of root penetration is a practical working rule of when a material is not considered to be media for plant growth. The depth to one of the following contacts is a criterion for defining soils in *Soil Taxonomy*.

Densic contact: Boundary between soil and underlying densic materials.

Lithic contact: Boundary between soil and underlying materials that is in a strongly cemented or more cemented rupture-resistance class. The underlying material is so hard and consolidated that digging with hand tools is impractical.

Paralithic contact: Boundary between soil and underlying paralithic materials. When moist the paralithic material can be excavated with hand tools.

Petroferric contact: Boundary between soil and an indurated layer of iron-cemented material. The iron-cemented material contains little or no organic matter. Digging with hand tools is impractical even when the material is moist.

Soil Composition and Characterization

Experienced pedologists can read a great deal from the morphology of a soil that they see in the field. However, modern soil classification systems place a great deal of reliance on quantitative physical and chemical data determined by laboratory measurements. The number of measurements that can be made on soil material is infinite. The selection of soil properties used in classification is based on the number of other important properties that co-vary with changes in the property under consideration. Analytical methods are selected that offer the best means of measuring compositional properties of greatest value in interpreting the soil properties for various uses.

Development of Methods and Criteria for Acquisition of Data. New methods of soil analysis are constantly developed. Before the 1930s, total or elemental analysis of the soil was widely practiced. Weight percentages of each element were reported in the oxide form. These weight percentages were then divided by the molecular weight of each compound to give molecular percentages. Ratios of the molecular percent values (called "molecular ratios") were then calculated for each of the soil horizons to determine losses and gains and thus infer the nature of the soil-forming processes. For example, molecular ratios of the silica (SiO_2) to the sesquioxides (oxides of iron and aluminum, Fe_2O_3 and Al_2O_3) were computed to determine if the latter two were accumulating with respect to the silica, an indication of a conceptual soil forming process called "laterization." Ratios of alkali and alkaline earth species (expressed as molecular oxides) to aluminum (Al_2O_3) and aluminum (Al_2O_3) plus silica (SiO_2) were used indicate leaching (Jenny 1941).

An example of the data collected under this system and of the types of interpretations that were made from them can be found in the 1935 *Atlas of American Agriculture* (Marbut 1935). Total elemental analysis data for the most important soils of the United States are tabulated by soil horizon, and molecular ratios carefully calculated. Total elemental analysis is now seldom used but is still very useful for some purposes.

In the 1930s came the realization that most soil clays were indeed crystalline and that the active or available portion of a nutrient element was more important from an agronomic point of view than the total amount present. In the 1940s came the development of new instruments and new procedures. These included X-ray diffraction units, colorimeters, flame photometers, spectrophotometers and new ideas for extracting solutions that removed the "active" portion of the nutrient elements present. These developments brought about rapid changes in the methods used for characterizing and classifying soils. New technologies continue to be developed that increase the kinds of analyses that can be performed on soil materials.

Collection of Soil Samples for Laboratory Analysis. Like the old recipe for rabbit stew that begins "first get the rabbit," the laboratory procedure for determining soil composition begins with directions for obtaining the soil samples. Selection of the pedons for sampling is a very important procedure, well worth taking a great deal of extra care. The time, effort, and expense invested in laboratory analyses of the profile samples will be wasted if the samples are not representative of the soil. The data resulting from the analyses likely will be used by a number of people for different purposes. These people will be misled and their interpretations and extrapolations erroneous if the samples are not representative, both as to appropriate pedon site and horizon depths. Soil samples should be taken from a pedon determined by field studies and observations to be truly representative of the area under study.

Soil samples are taken for a wide range of objectives. No single protocol is satisfactory for all purposes. However, if data generated by examination and analysis of the soil sample are to relate to the soil from which it was extracted, the location has to be recorded. A soil sample without a complete documentation of the location is merely a "hand full of dirt." Spatial locations by global positioning systems (GPS), legal property designations, and "metes and bounds" criteria are satisfactory for most sampling objectives. Documentation of depth(s) sampled is critical for most objectives. By convention depths are measured from the top of the soil surface after loose leaves and other undecomposed, identifiable plant material has been removed. If the loose organic materials are to be sampled their depths are measured above the soil surface.

The simplest protocol is to sample by defined depths from the surface. This is appropriate for some purposes, but sampling within horizon boundaries provides greater uniformity and avoids mixing obviously contrasting materials that may occur at the same depth if a distinct horizon boundary is present. Horizon samples also are better suited for making comparisons to other profiles of the same class or taxon of soil.

Sampling a complete soil profile or pedon is necessary to assess the behavior of that soil individual, always remembering that one sample site is only one specimen of an individual class and may differ in many respects from other sites of the same kind of soil. It is like sampling one tree among thousands of trees of the same species. A profile description is the first step in sampling a soil for classification. After the horizons within the profile are identified and marked on the pit face it is time to decide what volumes of material are to be sampled.

Should all of a horizon depth or only the midsection of the horizon be sampled? To answer this question, consider first how the data will be presented and used. For horizons with clear or abrupt horizon boundaries, a sample that mixes the entire horizon is often the best. Where horizon boundaries are gradual or diffuse a sampling of only the mid section of the horizon may be satisfactory. If only a small midsection of adjacent horizons is sampled, the transitional boundary between them can be inferred as intermediate between measured properties of the sampled depths. One pragmatic aspect to consider when selecting sampling depths is that two samples can be mixed, either physically in the laboratory before analysis or mathematically after analyses are completed, but once a volume of soil has been mixed in the field it cannot be separated.

Another important consideration in selection of sampling depths for pedon classification is the specific criteria of *Soil Taxonomy* or other classification system used to identify that pedon. It is prudent to consider diagnostic horizon and specific taxa criteria that have a potential to be critical in the correct classification of the soil being sampled. Criteria such as the clay content increase within a vertical distance used to define the upper boundary of argillic and kandic horizons can be specifically addressed by selecting appropriate sampling depths. Criteria such as base saturation percentage at a specified depth or at a lithic or paralithic contact are best addressed by taking a specific sample from the profile.

Many of the details and techniques of soil sampling for genesis and characterization studies are presented in the *Soil Survey Laboratory Methods Manual* (USDA-NRCS 2004). We recommend careful reading of this reference before detailed soil profile sampling is undertaken. We emphasize the following points about soil pedon sampling:

1. Uncultivated and virgin sites should be sought out for sampling, unless one of the major objectives is to characterize cultivated soils.

2. Road cuts and ditch banks generally are not satisfactory for sampling. The wetting-drying cycles, high degree of oxidation, plant root and animal activity, and contamination from dust or other atmospheric pollutants in these sites all tend to modify the soil structure and compositional properties to such an extent the samples may not be representative.
3. There is no satisfactory substitute for a well-located sampling pit, about 2-m long by 1- or 2-m wide and 2-m or more deep. Auger borings, probes, and tube samplings do not allow one to delineate and sample the various horizons satisfactorily as they occur in the pedon. There also is a high risk of introducing contamination as samples are lifted from a small hole.
4. One side of the pit, preferably the one that has the best lighting and where the profile description was made, should be sampled.
5. Although morphological boundaries are the primary determinants of sample depth, it is often desirable to collect a number of subsamples for a specific analysis such as determining the rate of change of a property with depth. Sampling a series of 5-cm-thick layers through the zone suspected of having the clay content increase specified for argillic or kandic horizons, regardless of horizon designation, is very desirable.
6. If a large number of undisturbed cores are to be taken and the soil is dry and hard, it is usually best to sample from the top downward. In this procedure, after a horizon is sampled, the remainder of the soil of that horizon in the sampling area is removed, exposing the next lower horizon for coring and bulk sampling. If taking cores is not a major consideration, then it is ordinarily best to sample from the bottom of the profile upward to avoid material falling onto unsampled horizons.
7. Bulk and core samples should be very carefully labeled, using tags or other identification both inside and outside the container.
8. For micromorphology study, blocks of undisturbed soil with orientation marked (as to which was the upper side, in place) can be collected and placed in containers that protect them from physical damage prior to thin-section preparation and microscopic examination.
9. Samples for microbial study may need to be refrigerated and/or protected from air.

Laboratory Determinations. From the many analytical methods used to analyze soil material a few have been selected to routinely characterize pedons for classification. Selection does not always imply that one method is superior to a competing method but by adopting specific methods results are more comparable among previously obtained data and the samples being studied. New methods are adopted as new types of equipment become available and sufficient comparisons establish correlation with existing data. A complete summary of all the methods available is beyond the intent of this book. Excellent references are available to detail analytical methods of soil analysis (Page et al. 1982; Klute 1986; Sparks 1996). The following sections discuss the chemical, physical, mineralogical and micromorphology determinations most often used in characterization of pedons for classification in *Soil Taxonomy*.

Cation Exchange Capacity (CEC). The capacity of a soil to attract and exchange species of positively charged ions (cations) in reversible chemical reactions within the soil solution is a quality important for soil fertility, waste disposal contamination and soil genesis studies. However, measurements of CEC are rather empirical, and several different analytical methods are used that yield different results (Chapman 1965; Coleman and Thomas 1967; Jackson 1958). Among the factors contributing to differing CEC values are the pH value at which the CEC determination is made and the species of ions used in the exchanging or displacing solutions. Ion species differ in ease of displacement, and certain species (potassium especially) may be trapped or fixed by some clay mineral species.

Despite these difficulties and problems, CEC determinations yield numbers that are valuable in evaluating the capacity of the soil to retain cations and in determining its clay mineral composition and general chemical reactivity. Three types of CEC determinations are commonly employed:

1. Ammonium saturation displacement method (CEC_7) conducted at pH 7.
2. Sum of cations method ($CEC_{8.2}$) performed at pH 8.2 in which the acid-generating hydrogen and aluminum (exchange acidity) are added to the basic cations (Ca^{2+}, Mg^{2+}, K^+, and Na^+) displaced by ammonium. This method of determining exchange acidity is buffered against pH fluctuation (Mehlich 1938), and in most soils yields CEC values higher than that from the ammonium saturation method. This is due to the fact that CEC values increase with higher pH in organic matter, 1:1 clay minerals, sesquioxides, and amorphous or short-range-order compounds. Soils rich in montmorillonite and other permanent-charge minerals do not display this feature.
3. In the ECEC method, the pH of the soil is not controlled, and the basic cations displaced by ammonium are added to the KCl-extractable aluminum.

Examples of how soil samples of different composition react to the different procedures are presented in Table 2.2.

The cation exchange capacity of organic matter is highly pH dependent, as illustrated by the two histic epipedon samples in Table 2.2. When the sample pH is 3.7 CEC greatly increases from the ECEC value to the CEC_7 value, but there is no significant difference between the ECEC value and CEC_7 value when the pH of the sample is 6.4. The CEC values of Oxisol samples in which the predominant clay mineral is kaolinite are greatly affected by the pH of the method, while samples from a Vertisol in which the predominant clay is montmorillonite are only slightly affected by the pH of the method. Because of this variation in CEC with pH, which is quite large in soils rich in kaolinite and hydrous oxides, it is imperative that the method used to determine CEC be specified. The conventions used to identify CEC methods differ. The $CEC_{8.2}$ method is identified as "sum of cations method" and the CEC_7 methods are identified as "NH_4OAc pH 7 method" in *Soil Taxonomy*. The ECEC method closely approximates methods referred to as "neutral salt" CEC or "permanent charge" CEC in some literature.

Table 2.2. Cation exchange capacity values of various soil materials (in cmol kg^{-1} soil) as determined by different methods

Soil Material	Organic Matter (%)	Soil pH 1:1 H_2O	ECEC Bases + Al	CEC_7 pH 7	$CEC_{8.2}$ pH 8.2
Umbric epipedon	14.0	4.7	14.5	36.7	48.0
Histic epipedon	90	3.7	16.0	103.8	nd
Histic epipedon	50	6.4	139.0	138.0	175.0
Oxisol A horizon	6	5.0	2.0	9.7	22.5
Oxic horizon	2	5.7	0.2	2.6	11.4
Vertisol A horizon	3.5	6.3	50.3	45.0	57.1
Vertisol B horizon	1.0	7.1	41.9	43.0	45.8

Cation exchange values of a whole soil sample are of limited value unless other features of the sample are known. The proportion of the CEC that is occupied by Ca^{2+}, Mg^{2+}, K^+, and Na^+ relative to that portion occupied by Al^{3+} and H^+ is of critical importance to plant root growth in acidic soils. If exchangeable Al^{3+} occupies more than approximately 60% of the ECEC, toxic levels of Al in soil solution result (Evans 1968; Nye et al. 1961). Some plants, especially legumes, are sensitive to as little as 20% of the ECEC occupied by Al. Thus, the exchangeable Al content is of importance and widely used in plant nutrition. Soil fertility specialists tend to prefer ECEC measurements because they reflect conditions at the soil pH and can be manipulated by adding calcium carbonate ("lime") to the soil. Soil classification tries to utilize criteria that are difficult to alter by common soil management practices such as lime applications; therefore greater use is made of CEC values that are determined at a specified pH value. In *Soil Taxonomy,* CEC values are used as the denominator to calculate the following percent base saturation values.

Percent Base Saturation ($CEC_{8.2}$). In *Soil Taxonomy* percent base saturation (PBS) by the "sum of cations" procedure at specified depths in a profile is used to differentiate Ultisols from Alfisols and as a criterion in various other categories of *Soil Taxonomy*. PBS by the sum of cations method is defined as:

$$\mathrm{PBS} = \frac{(\mathrm{Ca}^{2+} + \mathrm{Mg}^{2+} + \mathrm{K}^{+} + \mathrm{Na}^{+})\ \text{cmol/kg soil}}{\mathrm{CEC}_{8.2}\ \text{cmol/kg soil}} \times 100$$

Percent Base Saturation (CEC_7). Base saturation, calculated as above but using CEC_7 values in the denominator, is used to define the mollic and umbric epipedons and in various other places in *Soil Taxonomy*, where it is referred to as PBS by the "NH_4OAc method." In both cases, the basic cations (numerator) are extracted with 1 molar ammonium acetate at pH 7.

This seemingly contradiction in use of criteria is rational when one considers that epipedons are frequently cultivated, often have appreciable contents of organic matter, and receive lime to adjust their pH to near neutral. Deep in mineral soils organic matter contents are usually low, but the clay mineral species may differ among the soils in question and can best be compared at a specified high pH value. Also, a pragmatic consideration at the time *Soil Taxonomy* was being developed was the availability of data. More CEC_7 data were available for epipedons, but more $CEC_{8.2}$ data were available for the Ultisols in the southeastern part of the United States, where the Mehlich ($CEC_{8.2}$) method was more often used (Smith 1986).

Exchangeable Sodium Percentage (ESP). Soil materials disperse if they contain an appreciable amount of exchangeable sodium. One of the diagnostic features of natric horizons is an ESP of 15 or more. The ESP is calculated as:

$$\mathrm{ESP} = \frac{\mathrm{Na^+\ cmol/kg\ soil}}{\mathrm{CEC_7\ cmol/kg\ soil}} \times 100$$

Apparent CEC of the Clay. This calculated value is used to define oxic and kandic horizons (Kimble et al. 1993) and cation exchange activity classes at the family category in *Soil Taxonomy*. It is derived by dividing a CEC value of a whole soil sample by the percentage of clay in the sample:

$$\text{Apparent CEC of the clay (cmol/kg clay)} = \frac{\text{CEC cmol/kg soil}}{(\text{\% clay in the soil sample/100})}$$

Such a calculation assumes that all of the negative charge present in the sample originates in the clay fraction. This is obviously not correct because of negatively charged sites on organic matter and some silt- and sand-sized particles, but it is a useful measurement, especially in subsoils with low organic matter content. Apparent ECEC values of clay in acid soils are little affected by organic matter, and both values of 12 cmol ECEC or less per kilogram of clay and 16 cmol CEC_7 or less per kilogram of clay must be satisfied to identify the kaolinite-dominated low activity of kandic and oxic horizons.

Clay Mineral Inferences from CEC. Inferences about clay mineral species present in the soil can be made from CEC values and the clay content of the soil sample. Clay minerals in soils have been determined (Grim 1968) to have the following ranges in CEC_7 (cmol kg^{-1}): kaolinite, 3 to 15; smectite group (including montmorillonite), 60 to 100; illite (clay mica), 10 to 40; vermiculite (noninterlayered), 100 to 150; chlorite and hydroxy-interlayered minerals (HIM), 10 to 40. Clay can be separated and CEC measured directly, but a reasonable estimate can be made by assuming all of the CEC of a soil sample results from the clay and calculating an apparent CEC value of the clay.

pH Measurements. Methods of pH measurements of soil samples in the laboratory are described in many references. There are several field kits available for pH determination in the field, and if proper care is taken some are very reliable. To determine pH values the soil sample must be moist or in suspension. The amount of water in the sample affects the pH value obtained. In general, pH measurements of mineral soils are made in 1:1 soil:water ratio (that is, equal weights of distilled water and soil stirred in a beaker). Measurements of pH in 1:2 soil:0.01 M $CaCl_2$ suspensions are used to define reaction classes in families and pH values of NaF suspensions, 1 g soil in 50 ml of 1 M NaF, is used to identify isotic soil families. For organic soil materials, measurements of pH are made in 0.01 M $CaCl_2$ suspensions.

Soil Taxonomy seldom uses the pH of 1:1 soil:water mixtures to define taxa. Nonetheless, these measurements are desirable for pedon descriptions. The following inferences about soil chemistry can be made from pH values.

1. pH values less than 3.5 if obtained after repeated wetting and drying of the sample indicate the presence of acid sulfates (Fleming and Alexander 1961; Moormann 1963).
2. pH values less than 4.5 indicate a significant amount of exchangeable hydrogen is probably present in addition to exchangeable aluminum.
3. pH values below 5.2 in mineral soil indicate aluminum toxicity problems for most crop species (Kamprath 1967).
4. pH values 5.2 to 6.5 indicate acidity that may affect acid sensitive crops such as alfalfa.
5. pH values 6.5 to 8 indicate the soil is essentially fully base saturated; little or no exchangeable Al is present; free $CaCO_3$ may be present.
6. pH values 8 to 8.3 indicate the soil is fully base saturated and free $CaCO_3$ is present.
7. pH values 8.3 to 10 indicate appreciable exchangeable Na, perhaps enough to qualify as a natric horizon. Free sodium carbonate may be present.
8. pH values above 10 indicate the soil is highly sodium saturated. Free sodium carbonate is probably present.

Organic Matter and Its Components. Total organic matter content of soil material is commonly measured indirectly by determining the organic carbon percentage and multiplying by 1.724. Broadbent (1965) and Allison (1965) point out the ratio of organic carbon to organic matter is variable in different soils, and multiplication by 1.8 to 2.0 (approximately 1.9) is more appropriate for surface soils. For this reason most soil science publications report organic carbon percentage, but the popular term "soil organic matter" prevails in many publications and the conversion factor of 1.724 is most often used.

Soil Taxonomy utilizes only organic carbon values as criteria. Organic carbon content is most commonly determined after passing dry soil material through a 2-mm sieve to remove plant residues. The organic carbon content is determined by either dry combustion (heating in a furnace and measuring CO_2 evolved) or wet combustion (measuring the degree of reduction of a strong oxidizing agent) (Allison 1965; Broadbent 1965). The dry combustion-CO_2 evolution technique is quite quantitative (Allison 1965) and considered to be the standard method. The wet combustion procedure commonly used is the Walkley-Black method, in which the soil is digested in an excess of chromic acid, with titration of the unused oxidant (Allison 1965). Since some assumptions and approximations are involved in this procedure, it is less accurate and precise than dry combustion. In addition, this method generates Cr-rich waste, which is potentially hazardous. As a result, this method is much less commonly used now than in the past.

Organic carbon content is used to differentiate organic soil material from mineral soil material. Organic carbon content is a key criterion to identify anthropic, melanic, mollic, plaggen, and umbric epipedons, and spodic materials. Organic carbon content per square meter of soil to a depth of 1 meter is used as a diagnostic criterion in certain taxa in *Soil Taxonomy*. Percent organic carbon values are multiplied by 10 to obtain g C kg^{-1} of soil; the bulk density of each horizon within 1 m of the surface is determined, and the weight of 1 m^3 of the soil is calculated to determine these values.

Acid Oxalate Extractions. Aluminum, iron, and silicon contents in organic and short-range-order (amorphous) compounds such as allophane, imogolite, and ferrihydrite are determined in soils extracted with 0.2 molar ammonium oxalate, buffered at pH 3.0. The extraction is made with a mechanical vacuum extractor in the dark (USDA-NRCS 2004). The oxalate anion complexes Fe and Al in the organic and short-range-order compounds, causing these compounds to dissolve (also liberating Si). The data are used as criteria to define andic and spodic materials, as criteria for some mineralogy classes at the family level, and to assess the crystallinity of iron oxyhydroxides when compared to results from extraction with dithionite-citrate discussed below. Extremely small amounts of amorphous minerals and certain organic compounds impart chemical and perhaps physical properties to soil material because of their high surface area and solubility. These materials are not detectable by X-ray diffraction methods. Oxalate extractions have proven useful in identifying their properties and relating them to behavior of the soil material.

Dithionite-Citrate Extractions. Sodium dithionite is used to determine the content of secondary crystalline and short-range-order iron oxyhydroxides in soil (USDA-NRCS 2004). These compounds, which include hematite, goethite, lepidocrocite, maghemite, and ferrihydrite, are often referred to as “pedogenic iron” in that they represent secondary ferric-iron-containing compounds that formed primarily by the weathering of ferrous-iron-containing primary minerals. The ferric iron is reduced to ferrous iron by the sulfur in the dithionite, thereby making the iron soluble. The ferrous iron is complexed by citrate to keep the iron in solution for analysis. The crystallinity of the pedogenic iron is assessed by expressing the oxalate extraction results (Feo) and the dithionite-citrate extraction results (Fed) as a ratio (Feo/Fed). Ratios near 1 indicate low crystallinity of the iron compounds, often an indicator of weakly weathered soils, whereas ratios near 0 indicate highly crystalline iron oxyhydroxides in more strongly weathered soils. Dithionite extractable iron content is used to identify some mineralogy classes at the family level in *Soil Taxonomy*.

Phosphate Retention. Some soils, especially those with significant amounts of short-range-order minerals, immobilize large quantities of P and significantly reduce the effectiveness of fertilizer applications. This appears to be related to iron and aluminum contents as determined in oxalate extracts. A more direct method of evaluation P retention is to add 1000 g kg^{-1} P solution to a 5 g sample of soil, shake for 24 hours and determine how much P remains in solution (USDA-NRCS 2004). Values obtained by this method are used to define andic soil materials.

Saturated Paste Extract Electrical Conductivity. Appreciable accumulations of soluble salts are present in certain soils. Soluble salt content can be estimated by measuring the electrical conductivity of the extract from a paste made by saturating a soil with water. The saturated soil paste is vacuum filtered to obtain the extract (U.S. Salinity Laboratory Staff 1954). The amount of soluble salts in the soil is estimated by measuring the extract electrical conductivity with a conductivity meter. The results are expressed in units of decisiemens per meter (dS m^{-1}) at 25°C. A saturation extract conductivity of 30 dS m^{-1} at 25°C or greater is a diagnostic criterion for salic horizons.

Particle-Size Analysis. Samples are passed through a sieve to remove particles larger than 2 mm prior to particle-size analysis. The fraction <2 mm in diameter is the fine-earth fraction; the fraction ≥2 mm is referred to as "rock fragments." Particle-size analysis procedures have two aspects: (1) dispersion of the fine-earth fraction, and (2) fractionation into particle size classes. Most soil samples are dispersed by shaking or stirring the soil sample with a sodium hexametaphosphate solution. Determination of the proportions of the soil separates, sand, silt, and clay ordinarily is by a sedimentation-hydrometer procedure or by a sedimentation-pipette method, plus sieving to determine the sand content. A nest of sieves is used when sand subfraction percentages are to be measured (Day 1965; USDA-NRCS 2004).

The sedimentation-pipette procedure has been the method of choice. The hydrometer method is very acceptable when properly conducted. An older version of the hydrometer method that called for one hydrometer reading after 2 hours to determine clay content was often utilized in introductory laboratory courses and led to erroneous results. It normally takes from 6 to 8 hours, depending on the clay content, to properly measure clay content by the hydrometer method.

Problems encountered with dispersion are common when soil materials of contrasting mineral composition are analyzed. Carbonate, organic carbon, and/or iron may need to be removed prior to particle-size determination (USDA-NRCS 2004). Bartoli et al. (1991) have examined dispersion problems in strongly aggregated oxic horizon samples and recommend Na-resin dispersant and controlled ultrasonic or vigorous mechanical treatments. For such soils the percent water retained at 1,500 kPa times 3 closely approximates the percent clay (SMSS-USDA-SCS 1986). However, if samples contain appreciable halloysite or amorphous gels, the relationship is not valid.

For studies of clay movement and argillic horizon identification it is very useful to calculate ratios of the fine clay (less 0.2 µm diameter) to the coarse clay (0.2–2 µm diameter). This is based on the hypothesis that fine clay is more mobile in the illuviation process than is the coarse clay. Fractionation of clay-sized particles requires use of centrifugation (Jackson 1956; USDA-NRCS 2004).

Particle size analysis data are required for several criteria in *Soil Taxonomy*. In addition to particle-size family classification (Figure 2.2B), clay contents are needed to calculate the apparent CEC of the clay used in argillic, kandic, and oxic horizon identification.

Bulk Density. Bulk density of a soil is the ratio of the mass of the oven-dry soil to the volume of the soil. Soil bulk density is ordinarily expressed in units of g cm^{-3} or Mg m^{-3}. Bulk density can be estimated from soil porosity if the specific gravity of mineral particles is known. The average particle density of most soil mineral particles is close to that of quartz, 2.65 g cm^{-3}. Thus, a soil that is 50% pore space and 50% mineral particles has a bulk density of about 1.33 g cm^{-3}.

Some soils, through a combination of pore space, low specific gravity of the mineral particles, and high organic matter content, have low bulk density. A bulk density of 0.90 g cm^{-3} or less is a criterion to identify andic soil materials. Organic soil materials also have low bulk density, in some cases less than 0.1 g cm^{-3}.

Bulk density values are useful in evaluating chemical associations with plant roots that explore a volume of soil rather than a weight of soil. Any component measured on a weight basis of soil is less concentrated per unit volume of soil in a low bulk density soil than in a high bulk density soil. For example, greater soil organic matter content decreases the bulk density of soil but does not influence volumetric available water-holding capacity in coarse-textured soils and decreases it in medium- and fine-textured soil (Bauer and Black 1992).

Plant roots have difficulty entering soils with high bulk density. Soil texture affects the soil bulk density above which root penetration is restricted. In general, higher bulk density values are needed to stop roots in sands than in clays, but pore continuity and geometry are also important to root penetration (Daddow and Warrinton 1983).

Bulk density measurements require that the weight and volume of a soil sample be obtained without disturbing the arrangement of soil material. Methods of obtaining bulk density data are the core, plastic- or paraffin-coated clod, excavation, and gamma radiation densitometry techniques (Blake 1965). The gamma technique requires sophisticated equipment, and the excavation method removes soil and measures the volume excavated. The core method forces a cylinder of known volume into the soil. A segment of the core is removed and a given length measured to obtain a volume. The sample is then weighed, dried to oven dryness, and weighed again. The clod method requires that an internally undisturbed volume of soil, about the size of a fist, be gently removed from the horizon, contained in a fine net, dipped in resin to water-proof the clod, and allowed to dry before placing the sample in a container for transport to the laboratory. The volume of the clod is determined by weighing the resin-coated, waterproof clod in air and submerged in water (Brasher et al. 1966; USDA-NRCS 2004). The clod method is useful in collecting data from soil horizons too dense or brittle to sample with a coring device, and also allows calculation of shrink-swell potential or linear extensibility from the same clod specimens (Grossman et al. 1968). However, it should be noted that bulk density values from a coated-clod method ordinarily are relatively high compared to core or excavation methods because inter-ped void space is not measured.

Soil Moisture Parameters. Oven dryness, obtained by heating to 105–110°C, is the basis for most laboratory analyses. Many soil properties are affected by this degree of drying, so most analyses are made on air-dry soil materials, and the oven-dry weight is calculated from the water content of the air-dry soil material.

The percent (volume basis) of water retained at 1,500 kPa approximates the "permanent wilting point" of the soil. At this water potential, most plants are unable to obtain enough water from the soil to maintain turgor. Moisture content less than this value is considered "dry" for classification in *Soil Taxonomy*. Water retention at 1,500 kPa is chiefly controlled by the specific surface area in soil material (Richards 1965) and can be compared to measured clay contents. The ratio of 1,500 kPa water retention to measured clay content is about 0.4. If the ratio is either 0.25 or less or 0.6 or more, poor dispersion is suspected in the particle size analysis, and clay percent is estimated by the following formula: Clay % = 2.5(% water retained at 1500 kPa tension – % organic carbon). Multiplication of the 1500 kPa water percentage by a factor of 3 yields a good estimate of the clay percentage in oxic horizons (SMSS-SCS-USDA 1986). Water retention at 1500 kPa, of both undried and air-dried samples, is routinely used to classify families in Andisols and some Spodosols (Soil Survey Staff 1999).

Water retention at 10 kPa in sandy soils or 33 kPa in finer-textured soils is an estimate of the amount of water retained in the soil following free drainage against the force of gravity. This water content is often referred to as "field capacity" and is the value used to define "moist" soil in standard measurements used in *Soil Taxonomy*. The "water retention difference" (WRD) is the amount of water retained in a volume of soil between either 10 or 33 kPa, depending on sand content, and 1,500 kPa. The WRD value is also known as the "available water-holding capacity" (AWC) of a soil on a volume basis. Bulk density values are needed to determine the natural volume of soil used.

In general, tension table and pressure chamber techniques, using soil cores with field structure preserved, are employed for soil moisture retention measurements at low pressure (high potential) values (100–200 kPa). A pressure plate and membrane apparatus, using crushed soil samples, is ordinarily employed for the high-pressure (low potential) measurements, especially the 1,500 kPa limit (Richards 1965; USDA-NRCS 2004).

Shrink-Swell Capacity and Linear Extensibility. Most soils have the capacity to volumetrically expand when moist and contract when dry. This quality is quantified through the coefficient of linear extensibility (COLE) or potential volume change (PVC or swell index). The former is ordinarily used for pedological purposes, and the latter is more commonly used for on-site evaluation of possible highway routes or potential building sites (Franzmeier and Ross 1968; Cheng and Pettry 1993).

PVC values are obtained by measurement in a specially designed instrument involving a strain gauge (Henry and Dragoo 1965). COLE can be calculated from bulk density measurements at moist (10–30 kPa) and dry (1,500 kPa) moisture contents as

$$\text{COLE} = (\text{Bulk density dry}/\text{Bulk density moist})^{1/3} - 1$$

(Grossman et al. 1968; USDA-NRCS 2004). Schafer and Singer (1976) developed a method of measuring COLE using soil pastes.

COLE values are reported as cm cm^{-1},which when multiplied by the thickness in cm of the soil horizon in question equal linear extensibility of that layer. A *linear extensibility* value of 6 cm or more in the upper 100 cm of some soils is a criterion to identify Vertic subgroups.

Mineralogical Composition

For ease of understanding and discussion, it is convenient to separate soil minerals into two broad groups: (1) sand and silt size particles mostly inherited from the parent rock; and (2) clay-size silicate minerals, oxides and amorphous fine-grained compounds either inherited from the parent rock and formed in the soil. However, it should be noted that some "clay" minerals, mostly phyllosilicates, are often present in the silt- and sand-size fractions.

Sand and Silt Mineralogy. Most sand- and silt-size particles consist of a single mineral; however, small rock fragments in these fractions may contain more than one mineral. Microcrystalline aggregates of volcanic ash and chert are also mixtures of various minerals.

Sand-sized minerals and aggregates are best identified and percentages established by use of a petrographic microscope, after appropriate sample preparation. It is most convenient to observe particles of fine sand, very fine sand, and coarse silt size. Generally, it is necessary to identify and count 300 or more particles from one of the above size fractions to establish mineral composition with reasonable statistical assurance, but the number of particles counted must increase as the number of mineral species present increases. X-ray diffraction and infrared techniques also provide useful information for the characterization of these coarser fractions, but these methods lack the precision that can be obtained by use of optical mineralogy methods with the use of a petrographic microscope. Petrographic studies are quite time consuming and tedious—one reason that comprehensive soil mineralogical data are quite sparse. Procedures of sample preparation suitable for soil petrographic analysis are presented in standard references and texts (e.g., Brewer 1964; Milner 1962) and are summarized in the *Soil Survey Laboratory Methods Manual* (USDA-NRCS 2004).

Some uses and inferences from sand and silt mineralogy data obtained by petrographic techniques follow:

1. Evaluate uniformity of parent material for the various horizons of a profile.
2. Determine the presence of lithologic discontinuities (based on mineral species differences from horizon to horizon).
3. Evaluate soil fertility reserves based on content of weatherable minerals that release plant nutrients as they weather.
4. Evaluate the degree of weathering, based on ratio of weatherable to nonweatherable or resistant minerals.
5. Predict the course of clay mineral formation and soil development based on types of sand and silt-sized minerals present.
6. Establish relationships of a soil to geologic material from the presence of certain types of primary minerals or aggregates.

The presence of less than 10% weatherable minerals in the 50–200 µm (very fine sand and fine sand) particle size range is a diagnostic criterion for Oxisols in *Soil Taxonomy*.

Clay Minerals and Short-Range Order Minerals. Layer aluminosilicate clays, kaolinites, smectites, micas, and vermiculites constitute the main portion of the clay fraction of most soils. Iron and aluminum oxides are also present in significant amounts in certain soils. Short-range-order compounds, that is, minerals of variable chemical composition but with a predominance of Si-O-Al bonds, such as allophane and imogolite, are especially abundant in soils formed from volcanic ash. Structure and composition of clay minerals are described in a number of texts and reference books. These materials are best determined by a combination of X-ray diffraction, thermal analyses (differential thermal analysis, differential scanning calorimetry, thermogravimetric analysis), specific surface area measurement, and electron microscopic techniques. Methods for preparing soil samples for analysis are described in *Mineral in Soil Environments* (Dixon and Weed 1989).

Even with the techniques and instrumentation available, quantification of clay mineral species amounts in soil samples remains difficult. This is because most clay mineral species in soils do not behave in the well-defined manner of standard reference clay minerals. Clay minerals in soil often have variable chemical composition and poor crystallinity. In soils, clay mineral particles may be partially coated with oxides and/or organic compounds, and individual particles may be partially truncated. Methods of sample preparation for clay examination are empirical and not equally effective on all soils.

Formation and alteration of clay mineral species in soil are important considerations in soil genesis research. Although studies of clay minerals in soil are informative and a vital aspect of soil science, the objective of soil classification is to quantify the properties of soils. Therefore, the properties that clays impart to soil are more frequently measured and used in *Soil Taxonomy* than are the properties of individual clay species. The methods employed to determine amounts of clay species in soil samples as required for mineralogy classes in the family category of *Soil Taxonomy* are X-ray diffraction (XRD), thermal analysis, petrographic analysis of silt and sand fractions, and dithionite-citrate and acid oxalate extracts for iron, aluminum, and silicon (USDA-NRCS 2004).

Soil Micromorphology

Soil micromorphology refers to the description, interpretation, and measurement of soil components and features smaller than what can be readily seen with the naked eye. "Soil micromorphology is a tool to extend macromorphology," according to Wilding and Flach (1985). Micromorphological studies add a visual form to the substances determined from other types of analysis (Kubiena 1964).

Many micromorphological studies require the use of thin sections and petrographic equipment. Micromorphological study begins in the field with careful observation and recording of the location of the feature in the soil profile and the aid of a hand lens. Preparation of thin sections for microscopic observation can use a variety of methods (Bourbeau and Berger 1947; Brewer 1964; Buol and Fadness 1961; Cent and Brewer 1971; Grossman 1964; Murphy 1986). Most procedures involve drying the sample, evacuating air from the pores, and filling them with some kind of resin. Drying is not desirable in soils of high organic matter content or those with expanding 2:1 lattice clays. To avoid this, freeze-drying techniques may be used. Also, acetone (Miedema et al. 1974) or high molecular weight polymers such as Carbowax 6000 can be used in wet samples (Mitchell 1956).

Once a thin section is prepared, optical microscope techniques are similar to those used by geologists and mineralogists. A practical method to quantify the observations is the point count method (Anderson and Binnie 1961; Murphy 1986). Other imaging technologies useful in micromorphological studies are presented in a special issue of *Geoderma* (Mermut and Norton 1992).

Terminology. Illustrations and terminology for describing soil micromorphology has been presented in several books (Brewer 1964, 1975; Bullock et al. 1985; Douglas and Thompson 1985). There is no rigid system for describing thin sections; however, procedures outlined by Bullock et al. (1985) and Stoops et al. (2010) are the most commonly used currently.

The overall arrangement and organization of particles and voids as viewed in thin section are referred to as the *soil fabric*. Some of the components of soil fabric described by Bullock et al. (1985) are outlined in Table 2.3.

A *basic component of soil fabric* is particle size. Individual particles larger than about 10 μm can be observed as *single mineral grains* in thin sections. The mineral type of each grain of this size can be identified by petrographic techniques and used in identification of parent material and lithologic discontinuities in soil profiles. As minerals weather, a structural outline, or *pseudomorph*, of the original mineral remains, and secondary clay minerals forming from that mineral may be observed in direct contact with that pseudomorph. Such observations are useful in identifying weathering and clay-forming processes.

Table 2.3. Micromorphology features

Fabric Feature	Components	Component Properties
Basic components of coarse material and fine material (<10 μm) (groundmass)	Single mineral grains	Mineral and/or rock type
	Mineral pseudomorphs	Type of alteration
	Masses of mineral material	Contrasting particle mixtures
	Masses of organic material	With or without cell structure
Microstructure	Peds	Natural units surrounded by pores
	Fragments	Units formed by disturbance and rupture
	Apedal	Massive or structureless
Pedofeatures	Textural pedofeatures	Coatings on void walls
	Depletion pedofeatures	Depletions around void walls
	Crystalline pedofeatures	Mineral crystals
	Amorphous pedofeatures	Isotropic masses
	Fabric pedofeatures	Nonisotropic masses
	Excrement pedofeatures	Animal excrement

Mineral particles too small for individual particle identification are identified as *masses of mineral material,* and their appearance is described and compared to surrounding material. *Masses of organic material* indicative of specific plants or animals are often seen in thin sections. If the cell structure is visible, identification of the organism may be possible, but often only structureless organic masses are present.

The structural units described in field morphology (Table 2.1) can also be observed in thin sections. *Peds* are natural structural units surrounded by pore space (Figure 2.5 and Figure 2.6).

Massive or structureless soil material is *apedal. Fragments* are small volumes of contrasting material present within a horizon most often resulting from physical disturbance by cultivation, animal activity, or uprooting of trees.

Some of the most interesting and informative features observed in thin sections are the *pedofeatures*. Pedofeatures are discrete fabric units recognizable from adjacent material as concentrations of one or more soil component. *Textural pedofeatures* often indicate the movement of clay into the subsoil. Coatings of clay on the walls of pore space, especially on ped surfaces a frequently called clay skins, clay films, or argillans (Figure 2.7, Figure 2.8) and are among the required properties of argillic and natric horizons.

A thin zone surrounding a pore may be the site where material is removed. These sites are known as *depletion pedofeatures*. It is common to see depletion of red color immediately around root channels, which indicates that iron oxides have been reduced and removed. Such features are commonly known as *redox depletions* (Vepraskas 1994). A decreased concentration of carbonate minerals around a pore indicates that carbonates have been removed. *Crystalline pedofeatures* are indicative of mineral formation within the soil. Calcite or gypsum crystals often form below carbonate or

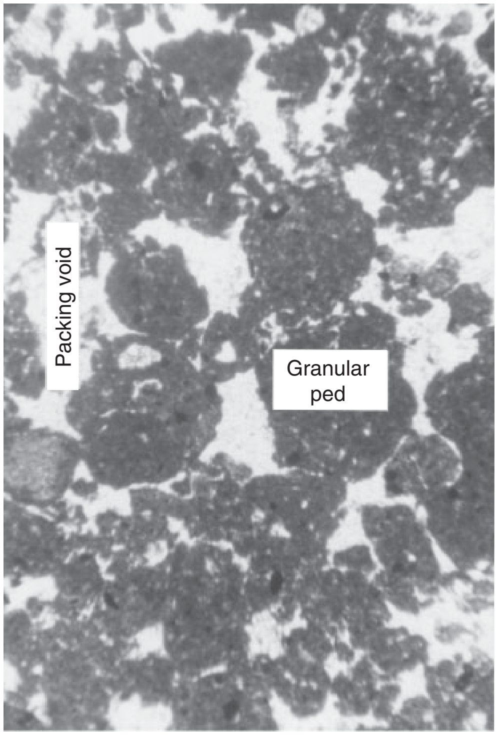

Figure 2.5. Granular peds in the oxic horizon of an Eutrustox pedon. (Photomicrograph with plane-polarized light; approx. X71)

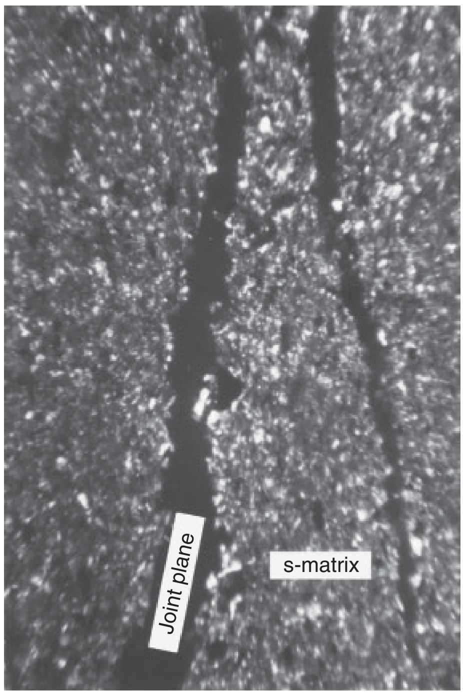

Figure 2.6. Platy structure near the surface of a Torrifluvent pedon. (Photomicrograph with crossed polarizers; approx. X71)

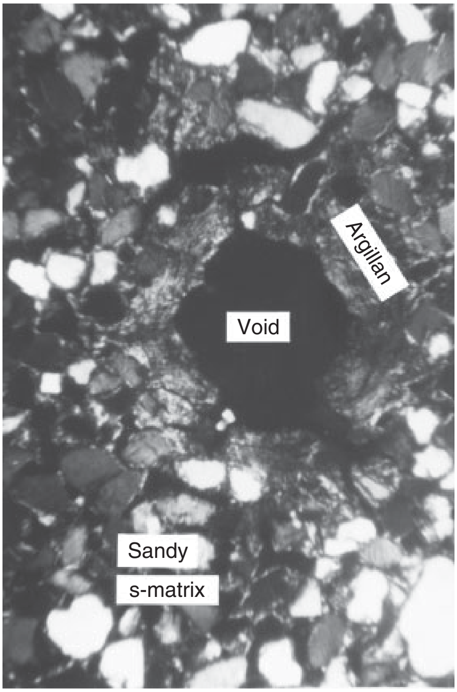

Figure 2.7. Argillan surrounding a void in the argillic horizon of a Haplustalf pedon. (Photomicrograph with crossed polarizers; approx. X71)

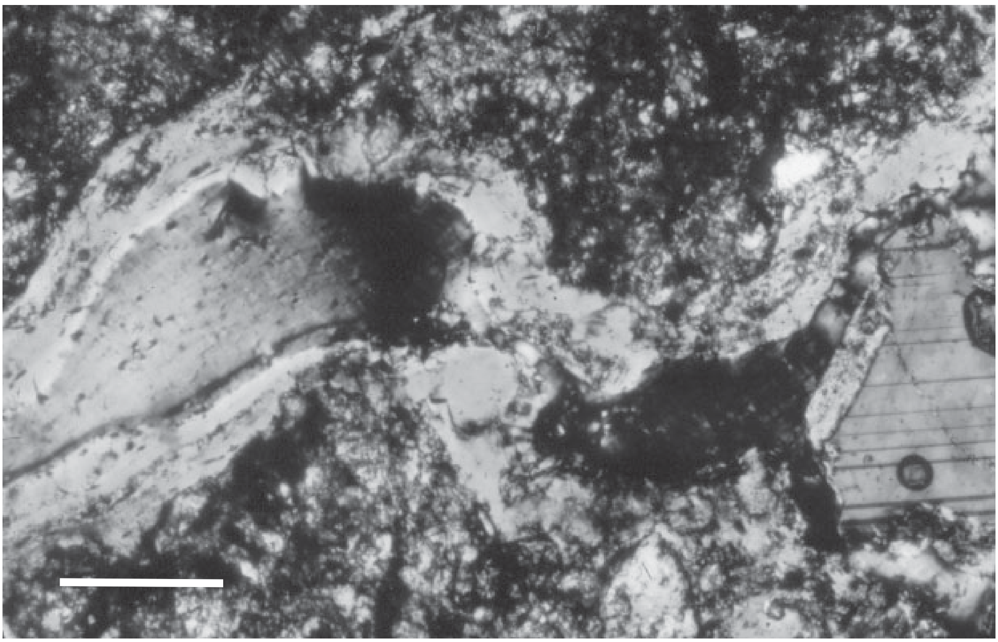

Figure 2.8. An argillan lining a pore in the argillic horizon of a California Palexeralf. Photomicrograph with crossed polarizers. The white bar represents about 250 µm. For color detail, please see color plate section.

gypsum depletions. *Amorphous pedofeatures* are coatings of oxide or organic material on void walls or concretions and nodules within peds. Oxide coatings indicate sites of oxidation in otherwise reduced soil and are commonly know as *redox concentrations* (Vepraskas 1994). *Fabric pedofeatures* are modifications of the internal fabric of a ped near the ped surface. The change in fabric is often diffuse and may result from the concentration of clay as shrink and swell action takes place upon wetting and drying in soils with appreciable contents of 2:1 clays. *Excrement pedofeatures* are animal excreta indicative of animal activity in the horizon examined.

Micromorphological studies also utilize scanning electron microscopes and electron probes that permit observation and nondestructive elemental analyses of volumes as small as 1 μm^3 (Cescas et al. 1968; Bisdom 1981; Goldstein et al. 1992).

Purpose Of Micromorphology Study. Most of the micropedological studies have been designed to aid in the understanding of soil genesis (Eswaran 1972), but they have also been used to study the durability of solidified and stabilized contaminants buried in soil (Klich et al. 1999). Micromorphology is also an important part of the study of soil-plant interactions (Melhuish 1968). Soileau et al. (1964) found that clay and iron oxide coatings on void walls reduced the uptake of potassium by plants. Khalifa and Buol (1969) found that natural clay skins acted to reduce P, K, and, to a lesser extent, N uptake by plants. Mücher and De Ploey (1977) employed micromorphology to determine the sorting of particles during raindrop splash and various intensities of runoff. Drees et al. (1994) studied soil structure resulting from no-till and conventional tillage.

Dynamic Soil Properties

Soil moisture and soil temperature are among the most important properties of a soil. An individual soil is fixed in space on the earth's surface, and how hot or cold and wet or dry a soil becomes determines how it responds as a medium for plant growth. No measures of soil morphology or composition have been found to adequately identify soil moisture and temperature dynamics. There has been much discussion about the validity and desirability of using soil temperature and moisture dynamics as taxonomic criteria (Smith 1973). Although temperature and moisture conditions in a soil are constantly changing, both can be measured and probabilities determined for each soil. Direct measurements of temperature and moisture dynamics require observations throughout several years, but reasonable estimates can be made from existing weather data. Probability measurements are not guarantees of what a given year or day will be like, but can be used to identify biological ecosystems and reflect the selection of crop species and use of drainage and irrigation systems by farmers. Seasonal dynamics of both soil temperature and moisture are reflected in traditional planting dates. *Soil Taxonomy* relies on estimations that can be made from long-term climatic data.

Soil Temperature Regime (STR). Mean annual soil temperature (MAST) helps determine where important agricultural crops can be grown in the United States.

In general, citrus crops are limited to MAST above 22°C, cotton is limited to MAST above 15°C, and corn for grain is limited to MAST above 8°C. As a rule of thumb in most of the United States, 1°C is added to mean annual air temperature (MAAT) to estimate the MAST. Van Wambeke (1981, 1982, 1985) added 2.5°C to MAAT to estimate MAST in tropical latitudes. Murtha and Williams (1986) found that in tropical Australia mean monthly soil temperature is best approximated by adding 4°C to mean monthly air temperatures in ustic soil moisture regimes and 2°C in udic soil moisture regimes. Where reliable air temperature data are not available, one-time measurements of soil temperature at depths greater than 10 m provide a satisfactory estimate of MAST (Smith et al. 1964).

To obtain direct measurements of soil temperature, temperature probes installed in the soil that record data digitally allow for nearly continuous measurements. Soil temperature can also be reasonably estimated by making one measurement at a depth of 50 cm near the middle of each month. By measuring the temperature at 50 cm, daily temperature fluctuations are avoided (Smith et al. 1964; Pikul and Allmaras 1984), but seasonal variations are reliably detected. The 50 cm depth is the standard depth for determining the STR in *Soil Taxonomy*.

Seasonality of soil temperature is of great significance to soil water balance and, hence, to ecosystem function and to agronomic practices. In the temperate latitudes, winter and summer seasons determine when food crops are planted and grown. In tropical latitudes and a few coastal areas in temperate latitudes where temperatures are modified by oceans, there is little difference in soil temperature throughout the year. The criterion used in *Soil Taxonomy* to differentiate soils in which seasonal temperature changes dictate agronomic crop scheduling is a difference of 6°C between the mean soil temperature of June, July, and August compared to the mean soil temperature of December, January, and February. Soils that have seasonal temperature fluctuation less than 6°C are identified as "*iso*" soil temperature regimes. In the absence of direct measurement, this limit can be approximated as 0.66 of the difference between the mean air temperatures of the designated months (Van Wambeke 1981, 1982, 1985). Murtha and Williams (1986) suggested that a difference of less than 5°C between the mean air temperatures of the above-named months is a better representation of an *iso*-STR than the Van Wambeke 0.66 factor. Cultivation of food crops in *iso*-STRs is generally limited by a MAST below 10°C.

Soil temperature during the summer months is very critical to soil use at high latitudes, where a short warm season with long daylight hours permits limited crop growth. Summer soil temperatures are colder when soils are insulated by an organic layer or have high heat capacity because of high water content. Details of the soil temperature regimes used in *Soil Taxonomy* are presented in Chapter 7.

Soil Moisture Regime (SMR). Too much water or too little water in the soil adversely affects plant growth. The amount of water in a soil is related to the climatic conditions over the soil and the position of a soil in the landscape. Moisture content in a soil can change rapidly. It is difficult to develop soil moisture criteria that can be used to characterize and classify soil, but failure to do so ignores perhaps the most significant function of soil in plant growth. A single observation is not sufficient to characterize the ability of a soil to provide adequate moisture for plants or identify conditions when too much water may be present. Current methods involve calculations of probability. This is quite realistic when one considers that climax vegetative ecosystems and agricultural stability result from compatibility with the soil over many years.

Methods to measure and characterize the soil moisture regime (SMR) probability are evolving in soil science. An early attempt (Soil Survey Staff 1975) used moisture conditions "present 6 out of 10 years." This part of the soil moisture regime definitions has been replaced with the term "*normal years*" (Soil Survey Staff 1999). Normal years are defined as years in which the mean annual precipitation is within plus or minus one standard deviation of the long-term (30 or more years) mean annual precipitation, and monthly precipitation is within plus or minus one standard deviation of the mean monthly precipitation for eight or more months during that year.

In the absence of direct measurements of soil moisture in normal years, estimates of soil moisture dynamics can be made from climatic data. In most cases, averages of long-term monthly rainfall and temperature are satisfactory for SMR characterization of soils on nearly level land, and worldwide SMR characterization data are available (Van Wambeke 1981, 1982, 1985). On landscape positions that experience extensive run-off or run-on, direct measurements of soil moisture content in normal years are the most reliable means for estimating SMR.

To define soil moisture regimes, three conditions of moisture content in a soil are considered. *Dry* soil is considered to be soil with moisture contents less than that moisture retained at 1,500 kPa. *Saturated* soil is defined as the condition when all the pore space is filled with water and water flows into an open hole bored into the soil. *Moist* soil has a moisture content between dry and saturated.

The following soil moisture regimes are defined and used in *Soil Taxonomy*.

Aquic Soil Moisture Regime. This is a regime of water saturation, oxygen depletion, and chemical reduction in a soil. Soils in which saturation of one or more horizon occurs in normal years are said to have *aquic conditions*. Aquic conditions are determined by direct measurements of saturated conditions during normal years, direct measurement of the redox potential of the soil, and observations of soil colors indicative of chemical reduction.

Saturation can be determined by systematically observing water table depths and duration at depth in an unlined auger hole through several normal years. In clayey soils with strong grades of blocky structure, open auger holes may fill with water when the soil is not saturated, so piezometers are used (Soil Survey Staff 1994).

Although it is desirable to obtain direct observations of saturation, this is not always possible. Evidence of saturation and chemical reduction is frequently inferred from the presence of *redoximorphic* features. Redoximorphic features are evidence that iron and/or manganese have moved under conditions of chemical reduction. Low chroma colors, usually chroma 2 or less, coupled with a moist value of 4 or more, generally indicate iron removal. Small volumes of low chroma color adjacent to higher chroma color are referred to as redox depletions. Adjacent small volumes of higher chroma, indicating accumulation of iron, are referred to as redox concentrations. Iron- and manganese-cemented concretions and nodules are indicative of aquic conditions. Saturation for only a few days may result in the formation of redoximorphic features. No exact length of time can be identified as necessary for redoximorphic features to form because the rate of microbial respiration that depletes the oxygen content in saturated soil and creates the reducing conditions varies with temperature and carbon availability.

The presence of reducing conditions can be determined by direct measurements of redox potentials with electrodes or by applying α, α'-dipyridyl, in neutral 1 M NH_4OAc solution, to a freshly broken, field-wet, clod surface and observing a reddish color indicative of the presence of ferrous iron.

Three types of saturation are defined: (1) *endosaturation*, in which all layers of the soil from the upper boundary of saturation to a depth of 2 m are saturated; (2) *episaturation*, in which the soil is saturated in one or more layers above 2 m, but some layer of the soil between the saturated layer(s) and 2 m is not saturated (for example, a perched water table); (3) *anthric saturation*, in which saturation in the upper layers of a soil caused by "puddling" the soil as in the controlled flooding of wetland rice or cranberry cultivation. Saturation, even for short periods, is a significant feature affecting many soil uses. Most soils with aquic conditions require artificial drainage before they are suitable for most cropping practices. However, aquic conditions in the lower part of the soil profile may be advantageous for crop growth due to the availability of water.

Aquic conditions are often related to landscape position and may be present in soils surrounded by soils with any of the following SMR. If aquic conditions are not present, the following procedures are used to identify the SMR.

Estimating Nonaquic Soil Moisture Regimes. Simple water balance calculations made from climatic data are used to estimate SMR for soils not strongly affected by endosaturation. The Thornthwaite (1948) method of estimating potential evapotranspiration is often used, but other reliable methods may be substituted. The water balance methods of Mather (1964; 1965) are employed to estimate soil moisture dynamics. A model developed by Newhall (1972) is very useful for determining SMRs.

For water balance calculations of SMRs, a moisture control section is defined in terms of water retention. The *moisture control section* is defined as that layer of soil below where 2.5 cm of water wets a dry soil within 24 hours and above where 7.5 cm of water wets a dry soil within 48 hours. Moisture contents less than permanent wilting percentage (1,500 kPa) are defined as dry. The moisture control section contains 5 cm of water available for evapotranspiration. The depth and thickness of moisture control sections vary depending on water retention characteristics and are approximately between 10 and 30 cm below the surface in fine-loamy, silty, and

Table 2.4. Example of a soil moisture regime calculation

Month	Ppt. (cm)	Pet. (cm)	Net (cm)	Store (cm)	Control Section
Jan.	10	0	10	7.5	M
Feb.	10	0	10	7.5	M
Mar.	10	2	8	7.5	M
Apr.	10	5	5	7.5	M
May	10	10	0	7.5	M
June	13	15	–2	5.5	M
July	13	20	–7	0.0	P
Aug.	10	18	–8	0.0	P
Sept.	10	5	5	5.0	P
Oct.	8	2	6	7.5	M
Nov.	8	0	8	7.5	M
Dec.	10	0	10	7.5	M
Year	122	77			

Notes: Ppt. = monthly precipitation: Pet. = potential evapotranspiration; Net = monthly Ppt. – monthly Pet.; Store = amount of H_2O in and above the control section at the end of each month. Within the control section M = control section is completely moist during the month; P = control section is partially moist during the month.

clayey soils, 20 and 60 cm in coarse-loamy soils, and 30 and 90 cm in sandy soils. In soils with a root-restricting layer, the moisture control section is limited to the amount of available water-holding capacity above that layer. Average or normal year climatic data are then used to calculate the probability of water being available in the moisture control section, which is the basis for defining SMRs.

An example of how average monthly precipitation (Ppt.) and potential evapotranspiration (Pet.) are used to calculate soil moisture regime is given in Table 2.4.

The amount of water available each month (Net) is calculated by subtracting Pet. from Ppt. for each month. The amount of water in and above the control section of the soil (Store) is most easily calculated starting in a month when the soil is either completely filled with water or completely dry. In the Table 2.4 example, (Net) is positive for both September and October and a sum of the 2 months exceeds 7.5 cm, so 7.5 cm, the maximum amount of water that can be stored above the bottom of the control section, is entered in the (Store) column for November. In December another 10 cm is added, but by definition the soil storage in and above the control section cannot exceed 7.5 cm of water, so at the end of December, the Store column is still 7.5 cm. Rainfall exceeds or equals potential evapotranspiration each successive month through May, so 7.5 cm can be entered in the Store column for each. During June Pet. exceeds Ppt. by 2 cm (Net = –2) decreasing the total stored water to 5.5 cm at the end of June. It is assumed that this water is taken from above the moisture control section, and therefore the control section is completely moist (M) at the end of June and remains so for a few days into July. The net deficit of 7.0 cm in July indicates that at the end of July there is zero storage. No water is added in August, so storage remains zero for that month. Ppt. exceeds Pet. by 5.0 cm of water in

September, and that value is reflected as storage at the end of that month. However, 2.5 cm of that water is retained above the control section, and the control section is only partially moist (P). In October, the soil above the bottom of the control section is filled to 7.5 cm and the control section to 5.0 cm by the addition of 6.0 cm of water and remains filled through November, confirming the selection of November as a starting point.

By convention one-half of the monthly rainfall is assumed to come on the 15th day of each month. Therefore, the control section is partially wetted in months during which rainfall exceeds 5 cm because one-half (>2.5 cm) of the monthly rainfall causes water to penetrate into the moisture control section. Thus, the moisture control section is not completely dry for the entire month, but may be considered dry for the first 15 days of that month. For this reason, the moisture control section in the Table 2.4 example is considered partially moist (P) during July and August, although the water balance at the end of each month is zero. This type of monthly water balance calculation is used to determine the following SMRs. The availability of moisture when it is too cold to grow most plants is of little value and, therefore, SMR definitions also include limits of soil temperature.

Most soils within a climatic area will have the same SMR, but exceptions can occur for two primary reasons. Soils that have a root restriction at a depth shallow enough to limit the amount of water within the moisture control section or that occur on slopes that enhance runoff are more arid than surrounding soils. It is common to have soils with an ustic SMR among udic SMR soils or soils with an aridic SMR among ustic or xeric SMR soils for one or both of these reasons. It is also possible that soils in valleys and landscape depressions receive run-on water and/or ground water seepage, producing a moister SMR than in surrounding soils.

Aridic, or Torric, Soil Moisture Regime. In normal years the moisture control section of these soils is both dry more than half the time when soil temperature are above 5°C at 50 cm depth and moist less than 90 consecutive days when soil temperatures are above 8°C at 50 cm depth. This temperature limit excludes aridic soil moisture regimes from soils with permafrost, isofrigid, and many cryic soil temperature regimes. Soils with an aridic, or torric, SMR do not have enough moisture to grow and mature most crops without irrigation. The terms "aridic" and "torric" are synonymous, but are used in different categories in *Soil Taxonomy*.

Perudic Soil Moisture Regime. In normal years precipitation exceeds evapotranspiration every month of the year (i.e., Net is positive in the SMR calculation). Harvesting and curing most grain crops is difficult because of the continuous humid condition.

Xeric Soil Moisture Regime. These soils are only in the temperate (non-iso-STR) areas, have MAST less than 22°C, and have dry summers and moist winters. In normal years, the moisture control section is moist in some part more than one-half of the days when the soil temperature is more than 5°C, dry in all parts more than 45 consecutive days during the four months following summer solstice, and moist in all parts more than 45 consecutive days during the four months following winter solstice. If winter temperatures are not severe, some growth is possible, but most crop growth takes place as temperatures warm in the spring by utilizing stored soil moisture.

Udic Soil Moisture Regime. In normal years, these soils are not dry in any part of the moisture control section as long as 90 cumulative days and do not meet the "dry in all parts for 45 consecutive days" criterion of the *xeric* SMR. In normal years, these soils have adequate moisture for growing crops any time the soil temperature is satisfactory for the crop, although drought conditions may occur in some years.

Ustic Soil Moisture Regime. In normal years, part of the moisture control section of these soils is dry for more than 90 cumulative days, but is moist in some part more than one-half of the days when the soil temperature is above 5°C at 50 cm. In soils with MAST of 22°C or higher or an *iso*-STR, if the moisture control section is moist in some part for more than 90 consecutive days each year. In soils that have a MAST less than 22°C and do not have an *iso*-STR, the moisture control section is moist in all parts at least 45 consecutive days during the four months following winter solstice, it is dry in all parts for less than 45 consecutive days during the four months following summer solstice. Soils with an ustic SMR and an *iso*-STR can grow one or two rain-fed crops per year, but lack of soil moisture can be expected during part of the year when soil temperatures are favorable for crop growth.

Soil Moisture Measurements in the Field. Saturation is usually determined by recording the water table depth in an open bore hole or piezometer. Clay soils present unique problems (Bouma et al. 1980). For unsaturated soil moisture control sections, a neutron probe, time-domain reflectometry (TDR), or gravimetric sampling is required. Complete moisture characteristic curves are needed to enable relating these measurements to the "moist" and "dry" definitions of the xeric, ustic, udic, and aridic SMRs. Direct measurement of moisture in the soil is desirable, but is expensive and time consuming. Several years are required to establish data from "normal" years, and the same soil may have widely different moisture contents because of different vegetative covers and management practices that alter infiltration and/or evapotranspiration rates. Obtaining quantitative criteria for classifying the moisture dynamics in soils is a worthy challenge for soil science.

Perspective

The terminology presented in this chapter is that used by *Soil Taxonomy*. History has taught us that we can expect changes in nomenclature as the state of knowledge changes. Proposals for nomenclature change are constantly being made. These proposals usually take the form of research papers in the scientific journals. Upon publication, these proposals are critically examined and tested by other soil scientists and with the passing of time either adopted or ignored. An increasing emphasis on quantification is evident throughout the development of nomenclature used to describe soil.

The development of nomenclature in many different countries has often led to misunderstandings (Dudal 1968b). Undoubtedly, communications in soil science would be aided if universal acceptance of a common nomenclature system could be achieved. Although this chapter has devoted coverage to only the criteria used in *Soil Taxonomy*, other systems of soil classification should be not ignored. (See Chapter 6.) For example, the World Reference Base (WRB) is a widely used system for soil correlation globally, and shares some terminology and concepts with *Soil Taxonomy*.

With training and experience, a person can learn much and interpret many qualities from careful observation and study of soils in the field. However, to acquire the quantitative information needed for classification and precise predictions of soil behavior, laboratory measurements to determine the composition of carefully collected soil samples are essential. Compositional data are most effective in enhancing our knowledge of soils when it can be properly incorporated into the morphology of soils in situ. The incorporation of quantitative criteria of soil temperature and moisture dynamics greatly enhances the practical applications of soil classification. Present definitions may appear crude and problematic, but ignoring these vital soil characteristics severely limits our ability to understand the interactions of soil with ecosystems and agriculture, and we are confident that research will improve present techniques.

Soil-forming Factors: Soil as a Component of Ecosystems

3

Soils have both inorganic and organic components and thus are influenced by the composition of the geologic material in which they form and the organic ecosystems that form in and over them. Both the inorganic and organic components respond to inputs of energy and water as dictated by the climatic conditions at the location where they form. Soils form as a thin veneer at the earth surface and thus develop characteristics in response to various topographic differences that alter regional inputs of climatic energy. Soils do not instantaneously form. The time span during which soils form can be considered intermediate between the relatively short life span of organisms and the relatively long periods of geologic ages. The morphological, mineral, and chemical composition of soil that we observe results from not only the present climatic, biologic, and topographic conditions, and underlying geologic material but are also a product of how these conditions have changed over time. As integral components of ecosystems soils and formation of the properties each individual soil possesses cannot be understood in isolation but requires an understanding of how the other components of terrestrial ecosystems influence soil formation.

In this chapter we will discuss how the soil-forming factors of parent material, climate, relief, organisms, and time affect soils formation. As reported by Simonson (1989, 1997), soil-forming factors were introduced to the United States by (Dokuchaev 1893) but received little attention except from Coffey (1912) until each of these factors was assessed as independent variables in a masterful treatise “Factors of Soil Formation” (Jenny 1941). In this chapter we will also expand that approach by considering the interactions of the factors in soil formation.

The Place of Soils in Ecosystems

A terrestrial ecosystem is a unit of landscape composed of both its *biotic* community (assemblage of organisms), and *abiotic* (nonliving) phases, all of which contribute to the domain that we call soil. The interpenetrating “spheres” of the planet, namely, the atmosphere, hydrosphere, biosphere, pedosphere, and upper lithosphere, constitute one vast ecosystem (Jenny 1980).

The upper boundary of soil is the interface with the air. The lower boundary that separates soil from “not–true-soil” (C and R horizons) is not easily defined. It may be

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

placed at the depth to which roots penetrate. However, the depth to which roots extend varies with type and stage of vegetative growth, and thus a more comprehensive lower boundary must consider media suitable for plant roots. As will be discussed later a depth of 2 meters is considered sufficient for soil classification in *Soil Taxonomy*. However, many significant alterations of geologic materials take place at far greater depths and must be evaluated to understand soil genesis (Douglas et al. 1967; Moody and Graham 1994; Graham et al. 1994).

Soil genesis deals with phenomena both at specific sites on land surfaces and with patterns of soil properties within entire landscapes. Operational limitations often confine observations and sampling to narrow spatial units, that is, soil profiles. However, features that occur on a regular spatial basis and spatially exotic features resulting from naturally occurring events like uprooting trees and human-induced activities of digging often disrupt soil features and must be carefully noted in studies of soil genesis.

The material in which soils form is the soil parent material (p). In many cases significant alterations of geologic material take place well below the volume now considered soil. Soils form at fixed geographic locations during long periods of time and thus are subjected to both past and present climate (cl) at that location. The inequalities in elevation and topographic position of a land surface comprise its relief (r) and spatially have considerable influence on the biologic components and distribution of water and temperature within regional climatic patterns. Plant, animal, and microbial organisms (ϕ) (the slash mark distinguishes this symbol from zero) populate soil in response to ambient climatic conditions both aboveground and in the soil, and their activities in turn affect the formation of specific soil properties. Sporadic alterations of organisms in response to events such as fire followed by a succession of species are common occurrences during the formation of soil at most sites.

The amount of time (t) that the material in a soil has been acquiring the properties and characteristics we observe is difficult to determine because the time interval involved in forming most soil characteristics greatly exceeds the capability of direct measurements. During soil formation, organic ecosystems may change in response to disruptions by fire, disease, and extreme climatic events. The vertical location of the soil surface is subject to change from erosion and/or deposition. Therefore, evaluation of time as a factor in soil formation is done by comparison of soil properties among soils formed on parent materials known to be of different geologic ages or among landscape positions subjected to different rates of erosion or deposition.

The relationships of the soil-forming factors can be expressed by the formula:

$$\text{Soil} = f\,(p, cl, r, \phi)\; t$$

Literally stated, the present properties of a soil result from a function of parent material, climate, relief, and organisms, over the period of time during which the soil formed. Although these conceptual considerations apply to all soils they do not take into consideration small increments of dust on all soils of the world (Syers et al. 1969;

Simonson 1995) and especially significant near active volcanoes and flood plains where there are significant incremental additions of new parent material to the soil surface without appreciable alteration of existing vegetation. Also, losses of material from dissolution and erosion are problematic components to be considered in the course of soil formation.

Yaalon (1975), Jenny (1980), and Smeck et al. (1983) provide thorough evaluations of conceptual models of pedogenesis and the inductive approach underlying them. A more recent review of both qualitative and quantitative pedological modeling is presented by Hoosebeek et al. (2000).

Parent Material

Early approaches to soil survey and soil classification were based on the geologic substratum (von Richthofen 1886; Thaer 1809, 1810, 1812). *Granite soils*, *glacial soils*, and *alluvial soils* were typical designations that revealed geologic emphasis. Such terms have some value in expressing soil properties and convey an early recognition of the relationship of soil properties to the geologic material in which soils form but obviously pertinent only to mineral soils that will be discussed in this section. The relationship of different types of organic materials to properties of organic soils is considered in Chapter 13 (Histosols).

The chemical, mineralogical, and physical composition of the parent material into which soil formation takes place is fundamental to understanding soil genesis. The solid mineral composition of soils is composed of *primary minerals*, crystallized at high temperature, and *secondary minerals*, mainly clay minerals formed from the decomposition of primary minerals at low temperatures during the process of soil formation (Churchman 2006).

Among the many parent materials from which soils form, many are deficient in or devoid of material necessary to support the operation of some soil-forming processes. Some are so resistant to change they are little affected by soil-forming processes. Simply stated in perhaps a homely analogy, a brick house cannot be built when only wood is available, and a wooden house cannot be built when only bricks are available.

In Table 3.1 data from C and Cr horizons of several parent materials in which soils form are presented. The data are from specific sites and do not express a range of properties that are present within the materials identified.

Siliceous Crystalline Rocks. Igneous and associated metamorphic rocks (collectively, commonly called crystalline rocks) may be grouped as follows: (1) felsic rock (*granite,* coarse-grained; *rhyolite,* fine-grained; and corresponding metamorphic gneisses and schists), (2) mafic rocks (*gabbro,* coarse-grained; *basalt,* fine-grained; and metamorphic equivalents, including greenstone); and (3) ultramafic rocks (peridotite and its metamorphic derivative, serpentinite) have very high magnesium contents but low aluminum, calcium, potassium, and sodium contents.

Table 3.1. Examples of C and Cr horizon composition of various parent materials

							Exchange Bases				
		Depth	Sand	Silt	Clay	pH	Ca	Mg	Na	K	$CaCO_3$
Material	Location	cm		%		H_2O		cmols Kg^{-1}			%
Granite Cr	Malaysia	1500–1600	70	22	8	4.9	0.02	0.07	–	0.05	0
Schist Cr	Malaysia	950–1150	28	60	12	4.6	0.02	0.04	0.02	0.8	0
Gabbro Cr	N. Carolina	165–203	92	6	2	6.8	2.7	1.6	0.1	Tr.	0
Till	Wisconsin	160–180	59	30	11	8.3	3.2	1.0	Tr.	0.1	24
Till	Minnesota	310–360	49	33	18	8.2	13.3	1.9	0.1	0.3	14
Till	New York	105–152	80	20	0	5.5	0.1	Tr.	0.6	Tr.	0
Till	New York	150–160	85	15	0	5.8	0.2	Tr.	0.6	Tr.	0
Outwash	Minnesota	142–177	94	4	2	8.4	2.7	0.8	0.2	0.1	2
Outwash	Maine	125–162	95	4	1	5.5	0	Tr.	Tr.	0	0
Lacustrine	Wisconsin	124–152	5	23	72	8.5	8.8	14.3	0.5	0.8	14
Loess	Iowa	172–192	3	71	26	7.0	14.3	8.2	0.2	0.6	Tr.
Volc. Ash	Ecuador	185–205	73	24	3	8.4	3.8	0.9	0.8	0.2	–
Volc. Ash	Chile	157–210	72	28	<1	6.0	4.7	2.6	–	Tr.	–
Volc. Ash	Chile	147–217	95	5	0	6.7	0.3	0.1	–	–	–

Sources: Iowa, Minnesota, and Wisconsin: USDA-SCS 1978a; 1978b; 1979, respectively: Malaysia: Department of Agriculture 1978. North Carolina: Rice, et al. 1985a. Chile: Beinroth et al. 1985a. Ecuador: Beinroth et al. 1985b.

Soils formed from mafic rocks are usually more fertile, with higher contents of calcium, magnesium, and phosphorus than soils formed from felsic rocks. They also usually contain more smectite clay and have a more reddish brown color. Soils formed in ultramafic rocks, particularly serpentinite (Wildman et al. 1968), often give rise to "barrens," with plant communities characterized by endemic plant species that have survived, despite the nutrient imbalance of excess Mg and inadequate Ca (Brooks 1987). Coarse-grained rocks tend to weather faster than fine-grained ones, and mafic rocks, faster than felsic ones. One or two million years may be required for conversion of solid rock to deep soil in situ (Pavich 1985). Quartzite, a metamorphic rock of pure quartz (SiO_2) rock, is essentially infertile and resistant to soil formation, and most soils formed are shallow.

Siliceous Crystalline (Felsic) Rocks. These will be further discussed under two lithologic groupings:

1. *Granite and granite gneiss*. These rocks average about 25% quartz, 65% feldspar (dominantly alkali feldspar), with lesser amounts of mica (muscovite and biotite), and small amounts of hornblende. Weathering patterns follow rock fabric, from isotropic in granite to banded in gneiss, but resulting soils are very similar. In well-leached environments, the soils formed are acid, have low contents of base cations, and low mineral nutrient reserves. Kaolinite and hydroxy interlayered minerals (HIM) are abundant clay minerals formed in surface and subsoil horizons.

   In humid climates (udic soil moisture regimes), considerable physical and chemical alteration of the parent rock takes place well below the depth usually considered as soil (Table 3.2).

   The C horizons to a depth of 182 cm in Table 3.2 have weak subangular blocky structure and discontinuous clay films indicating some influx of clay from above. Below 182 cm the clay appears to have formed from the restructuring of ions as the voids saturate during periods of water influx and become supersaturated as they dry forcing the dissolved ions to precipitate forming gibbsite and kaolinite (Calvert et al. 1980b). The dissolution of the rock is evidenced by the marked reduction of bulk density in the lower C horizons that retain their 'rock-like' structural form and conform to the definition of saprolite, that is, "soft, earthy, clay-rich, thoroughly decomposed rock formed in place by chemical weathering of igneous and metamorphic rocks" (Glossary of Geology 1972). Note that silica, calcium, magnesium, and sodium contents greatly decrease upward in the column while aluminum and titanium contents increase as the more soluble silicate minerals are removed by leaching. Enrichment of calcium and potassium in the A horizon may be attributed to biocycling by plants. Enhancement of silica content in the A horizon can be attributed to the greater sand content relative the greater clay content in the Bt horizon. The most significant aspect with respect to soil genesis is that potassium, calcium, and sodium contents are greatly reduced well below and prior to the formation of the A and B horizons of the solum.

Table 3.2. Bulk density and total elemental analysis (weight percentage) of a site in North Carolina

Horizon	Depth (cm)	Density g cm^{-3}	SiO_2 %	Al_2O_3 %	Fe_2O_3 %	TiO_2 %	CaO %	MgO %	Na_2O %	K_2O %	Clay %
A	0–18	0.95	79.2	13.0	2.6	3.8	0.1	0.08	0.23	2.21	24.7
Bt	18–67	1.26	60.1	23.8	5.6	2.6	Tr*	0.18	0.07	1.88	60.7
C1**	67–102	1.32	67.2	21.1	4.0	2.4	Tr*	0.06	0.08	2.52	31.8
C2**	102–137	1.39	65.1	21.7	4.1	2.3	Tr*	0.17	0.08	2.55	30.4
C3**	137–182	1.39	69.3	19.9	3.5	2.1	Tr*	0.07	0.09	2.82	23.3
C4 **	182–217	1.32	74.9	18.9	2.7	1.7	Tr*	0.10	0.09	2.99	15.5
C5**	402–437	1.37	64.8	20.7	3.7	1.8	Tr*	0.26	0.11	3.55	5.7
C6**	512–522	1.53	71.7	16.9	1.5	Tr*	Tr*	0.06	0.61	4.08	3.4
R1	1,200	2.56	79.4	10.8	2.5	Tr*	1.4	0.48	4.37	2.05	–
R2	1,200	2.64	78.5	11.0	2.5	Tr*	1.6	0.54	4.50	2.09	–

Source: Calvert et al. 1980a.
* Tr = Less than 1.0 percent.
** Horizons C1–C6 renamed from original paper. Cr (saprolite) designation was not available at time of publication but applicable to C5 and C6.

Soils formed in these materials also have a considerable reduction in saturated hydraulic conductivity in the upper C and B/C horizons compared to the overlying, more clay-rich, but strong blocky structured Bt horizons (Shoeneberger et al. 1995). This is attributed to formation of clay films in the upper C horizons (Vepraskas et al. 1996). Many felsic saprolites contain quartz veins that appear capable of rapidly conducting water. However, one study found that such apparent quartz vein channels did not conduct water significantly faster than surrounding saprolite due to infilling of clay and/or iron oxides (Williams and Vepraskas 1994). Although saturated hydraulic conductivity is slowed in the upper zones of saprolite, water does move vertically within the saprolite (Li et al. 1997). In part the pathways for water to move into granite and other igneous rock is aided by the formation of unloading joints as over burden is removed by erosion and the rock becomes less confined (Holzhausen 1989). Proximity to the land surface also subjects the rock to temperature-induced expansion and contraction, especially in temperate latitudes.

2. *Schists*. These are foliated (thin, platy) metamorphic rocks rich in mica or chlorite, with quartz and small amounts of weatherable minerals. Saprolite formed from schist tends to contain more silt than saprolite from granite, as do soils derived from it (Table 3.1). Potassium reserves are usually high. Mica and vermiculite clays tend to predominate, but kaolinite and HIM contents tend to increase in the more acid soils. Weathering of aluminosilicates accounts for prominence of extractable aluminum in these acid soils. Smectite is present in soils with a marked dry season. Soils formed in chloritic schists are commonly clayey, plastic, smectitic, and possibly rich in magnesium.

Dark-Colored Ferromagnesian (Mafic) Rocks. These are the andesites, diorites, basalts, gabbro, and hornblende gneisses, all rich in iron- and magnesium-bearing minerals and calcic plagioclase feldspars that weather rapidly, yielding a good deal of clay and free iron. As long as they persist, these minerals maintain a higher base status than soils formed from felsic rocks in humid climates (Table 3.1). The near absence of quartz accounts for relatively low sand content in developed soils. Surface soils are most often loams and clay loams. Extractable aluminum levels are low compared to soils formed from siliceous rock. Kaolinite predominates in the dark red clays in well-drained soils, but smectite is more common in the fine clay fractions (Rice et al. 1985a). Iron oxides appear mainly as small discrete clusters in electron micrographs (Rice et al. 1985b).

Ultramafic Rocks. These rocks, mainly peridotite and serpentinite composed almost entirely of ferromagnesian minerals without feldspars or quartz occur as sills, dikes, and other small intrusive masses. Soils formed in ultramafic rocks have high exchangeable magnesium contents, low exchangeable Ca/Mg ratios, and high contents of Cr, Mn, Fe, Co, and Ni. The chemical composition of these soils, especially the low exchangeable Ca/Mg ratio appears responsible for unique communities of plant species. Woody vegetation is relatively sparse and/or stunted when compared to surrounding areas. These vegetative conditions are commonly referred to as “serpentine barrens” (Alexander 2009).

Sediments and Sedimentary Rocks. Blatt and Jones (1975) determined that sedimentary rocks underlie soils in 67% of the land areas of earth. All sediments and sedimentary rocks have one thing in common that differs from crystalline rocks. All minerals contained in sediments have been exposed to weathering and mineral decomposition at or near the land surface and/or during transport. Thus to some degree they have acquired mineral and chemical properties that represent conditions prior to deposition that cannot be attributed to processes in the soils that are currently formed in these sediments.

1. *Sandstones (quartzose sandstones and orthoquartzites)* contain more than 50 percent sand-size particles predominantly of quartz, the rest being impurities such as feldspar and mica, and cementing material usually silica, iron, or carbonates. In well-leached environments, most soils derived from these rocks are coarse-textured and deep. Except where the sandstone is cemented by carbonates most soils are acid, and low in base status and nutrient reserves. If feldspar content exceeds 25 percent, the rock is arkosic sandstone, and the derived soils may be clayey.
2. *Limestones and dolomites* are by definition composed of more than 50 percent carbonates, the remainder being “contaminants” deposited on the sea bottom, including silt, clay, and particles of quartz and iron. Some of the quartz was deposited as silica gel that later dehydrated to chert and flint, which are abundant in residuum from some carbonate rocks as in the Flint Hills of the Great Plains

of the United States. Although limestone forms in sea floors, it is a common component of mountains such as the Alps, Himalayas, and Andes, having been thrust upward from the sea by tectonic movements at the margins of tectonic plates. Soils formed over carbonate rocks commonly have neutral to slightly alkaline pH values and are rich in calcium and magnesium, except where surface horizons are leached, reddish in color, and clayey in texture. In many areas, the soils formed are an accumulation of insoluble residues derived from contaminants that were present in the limestone. In other areas, most of the clay in the red Bt and BC horizons that form over the carbonate rock is apparently eolian material that has been translocated downward in the profile (Frolking 1978). Where an area of limestone that is nearly pure $CaCO_3$ is exposed soils tend to be shallow.

3. *Shales* are laminated or fissile (thinly bedded), somewhat indurated or hardened rocks called clay shale, claystone, or mudstone, if predominantly clayey, or siltstone if silty. Layer silicates, feldspars, and quartz are the predominant minerals. Soils formed in shale tend to be clayey, impermeable, little leached, and shallow. Base status and pH values are commonly high when carbonate is present in the shale, but acid soils form in acidic shale. Mica and smectite are common clay species in soils formed in neutral to alkaline shale, but kaolinite is more common in acid shale. Soils formed from siltstones are silty to loamy texture, with medium to high nutrient reserves.
4. *Unconsolidated sediments subjected to anoxic (reduced) conditions* underlie extensive plains along coasts undefended by cliffs. In the United States extensive areas of coastal plain sediments form a border along the Atlantic and Gulf coasts. The lowlands of the Amazon Basin extending inland for over 2,000 km are perhaps the most extensive example of such unconsolidated sediments. The sediments are largely of marine origin at lower elevations, and of alluvial deltaic origin farther inland. Many and perhaps most of these sediments have experienced several cycles of erosion and deposition prior to being deposited in their current locations. With few exceptions derived soils are acid, contain few weatherable minerals, and are kaolinitic with some hydroxy interlayered minerals, except for some marine clay sediments that are smectitic. Sediments formed under these conditions have been subjected to reducing conditions during which iron and manganese oxides have been reduced and removed and have low iron and manganese oxide contents (Table 3.3).
5. *Strongly weathered oxidized sediments* deeply cover the igneous and metamorphic rocks of continental shields in South America and Africa. This cover often includes layers of pediment gravel, hillwash, and stone lines (Fölster et al. 1971). (See Figure 3.1.)

In South America the Brazilian and Guiana shields have subdued landscapes resulting from recurrent episodes of weathering and erosion during humid climates of Cretaceous and early Cenozoic time that generated deep, weathered residues of quartz, iron and manganese oxides, and kaolinite and gibbsite clays (Orme 2007). Unlike sediments deposited in coastal plains and other reducing environments, these

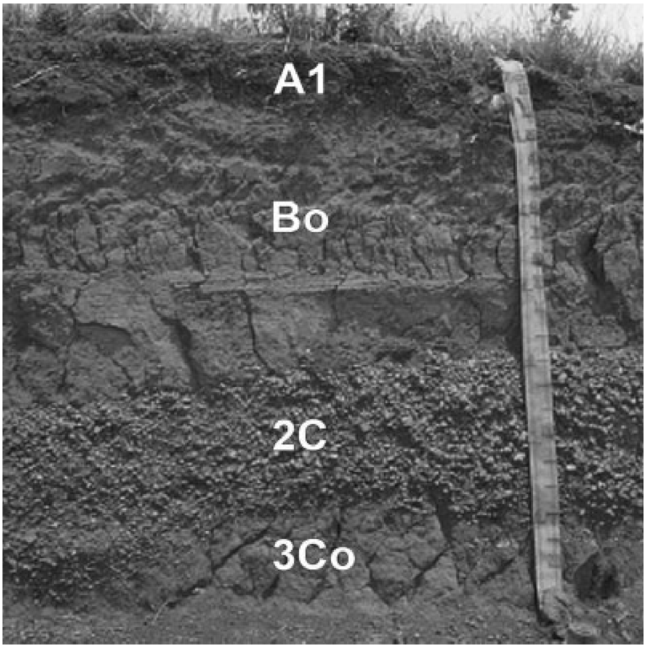

Figure 3.1. Profile of oxidized sediments in Federal District of Brazil. Note layer of ironstone and quartz gravel. Also note blocky structure appearance is due to road bank exposure. True structure of all horizons is granular. For color detail, please see color plate section.

Table 3.3. Examples of sediments in reducing coastal plain environment and sediments transported in oxidized environment

		Depth	Sand	Clay	DCB* Fe	Exch. Bases	pH
Horizon	Location	cm	%		%	cmols kg^{-1}	H_2O
C[1]	Florida	145–155	61	37	0.1	2.9	3.5
C[2]	North Carolina	270–345	70	27	1.2	–	–
C[3]	North Carolina	224–264	72	10	0.02	0.09	3.7
C[4]	Brazil (Basalt)	420–450	27	71	15.1	2.6	6.6
C[4]	Brazil (Conglomerate)	335–400	35	34	9.6	0.2	5.4
C[4]	Brazil (Reworked)	410–430	37	53	7.8	0.5	5.3

* Dithionite, Citrate, Bicarbonate Extractant.
Sources:
[1]USDA-SCS 1969.
[2]Daniels, R. B., E. E. Gamble, and W. H. Weaver. (Ceria 1970)
[3]FAO-UNESCO.1967.
[4]SMSS-USDA-SCS 1986.

sediments apparently have been little affected by reducing conditions. As a result most silicate minerals, except sand-sized quartz, have been weathered, but iron oxide contents are high, which favors aggregation of clay into granular peds (Table 3.3). These strongly weathered sediments range in texture for sandy to clayey, have low silt content, and are almost all confined to upland shield areas in the tropics. Soils formed on these sediments that have been through one to several weathering cycles are from 1 to several meters deep with solum features little changed from the parent material except increases in organic carbon content in surface horizons.

Glacial Deposits. Glacial deposits are unique among the many parent materials. They are the only parent materials that are physically pulverized into small particles and then abruptly deposited on the land surface in the zone of soil formation. These previously unweathered materials, except where sedimentary rocks have been incorporated become in situ sites of mineralogical alteration and release of ions in soils formed as the glaciers retreated. As a parent material, these physically pulverized and previously unweathered materials are unlike weathered saprolite materials formed over crystalline bedrock and/or sediments previously exposed to weathering at another site and/or during transport. Examples of various glacial deposits are presented in Table 3.1.

1. *Glacial till* is material that is directly deposited as glacial ice melted. Till deposits can be of any texture depending on the type of material the glacier encountered. In most cases tills contain a random distribution of angular rocks in a matrix of finer material. In the North American Midwest and much of Europe, loamy textures predominate in tills because the glaciers scoured limestone and shale. In these tills smectite and mica are often the dominant clay minerals, pH values are neutral to slightly alkaline, and base saturation percentages are high. In the northeastern states of the United States, and northern Europe, tills are sandy to sandy loam in texture, largely as a result of the passage of glaciers over granite and other acid igneous rocks. These tills, and the soils derived from them, have low pH values and base saturation percentage. The clay mineralogy includes mica and smectite with some kaolinite, vermiculite, and chlorite.
2. *Outwash deposits* consist of coarse-textured stratified deposits of sand and rounded gravel. These deposits resulted as the glacial ice seasonally melted and large rivers flowed forth. Most outwash deposits occur on broad nearly level landscapes. Soils formed directly in such deposits are usually sandy and have a high content of gravel.
3. *Loess* is primarily windblown silt-sized particles, with some fine sand. Loess blankets and is the parent material for soil formation in large areas in the North American Midwest, Eastern Europe, Central China, and many other places. The most extensive areas of loess are near the terminus of the continental glaciers and resulted as winds eroded silt from the seasonally dry and sparsely vegetated river beds and floodplains that resulted from the seasonal melt of glacial ice in these temperate latitudes. Some loess deposits have resulted from silt eroded from sparsely vegetated arid and semiarid areas. Although most geologists and soil scientists consider loess to be only wind-deposited material, water-laid silts, eroded from loess are also considered to be loess by Chinese scientists.

Table 3.4. Total major elemental composition (Wt. percentage) of loess samples (After äjvári et al. 2008)

SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5
70.71	0.71	11.74	3.75	0.07	2.15	6.67	1.68	2.22	0.14

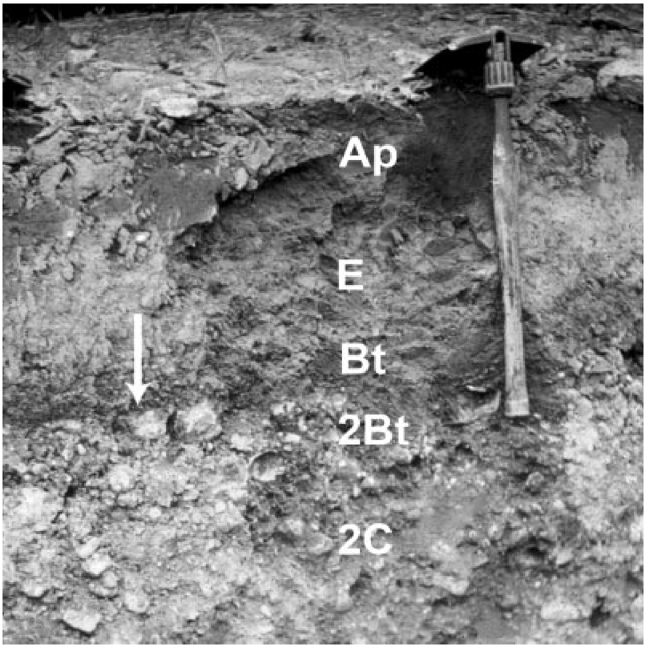

Figure 3.2. Soil profile developed in loess over till in Southern Wisconsin, USA. (Arrow points to contact between loess and till.) For color detail, please see color plate section.

Loess is typically composed of about 85 percent silt and 15 percent clay, with an abundance of weatherable minerals; a high base saturation percentage, often calcareous and contains clays of smectite (dominant), mica, and vermiculite species. Üjvári et al. (2008) averaged the elemental composition of 244 samples of loess from 11 loess areas around the world (Table 3.4). Although there is some deviation in values, there was remarkable agreement among all sites. Soils formed in such silt-rich material have high plant-available water-holding capacity, large nutrient reserves, and good physical conditions for plant growth.

In many glacial areas stone-free loess deposits of varying depths cover till and outwash deposits forming a lithologic discontinuity in the soils that formed. (See Figure 3.2.)

4. *Lacustrine deposits* are fine-textured sediments that accumulated in lakes that formed and persisted for many years as the continental glaciers were receding.

Lacustrine deposits consist of thin alternating layers, usually of silt- and clay-sized particles (varves) as turbulence in the lake water changed from wind turbulence in the summer to stillness when ice covered the lakes in winter. Soils formed in lacustrine deposits are often quite impermeable and in the saturated conditions created often have acquired and preserved organic matter residues thick enough that organic soils (Histosols) have formed.

Volcanic Ash. Although comprising only a small portion of the earth's land mass, these materials serve as the parent material for unique soil properties. The composition of volcanic ash is variable (Table 3.1). In contrast to the formation of feldspar, mica, and other primary minerals formed in the slow cooling of magma of igneous rocks deep in the earth's crust, volcanic ash cools rapidly as it is exhausted from volcanic vents, and glass, amorphous, or short-range-ordered particles are formed. Glassy or amorphous structures are more easily dissolved than structured minerals. During dissolution K^+, Mg^{2+} and Ca^{2+} ions are first released followed at a somewhat slower pace by release of Si and Al. In humid climates silicon and basic cations are leached while Al remains to form allophane-imogolite and Al-humus complexes in the soil (Shoji 2006).

The way in which volcanic ash material enters soil as periodic deposits on the surface of existing soil is unique. This combination of periodic additions of rapidly weathering minerals to the soil surface has given rise to fertile soils that in areas with favorable temperature and moisture conditions have long supported the growing of human food crops resulting in high human population densities in for example in Java, Japan, Central America, and the rift valley in east Africa. Most of the soils formed from volcanic ash are now classified as Andisols. They present unique analytical challenges for soil scientists as will be discussed in Chapter 9.

Lithosequences. Parent material sequences on a landscape are called lithosequences. The dependency of soil characteristics on the properties of the initial materials in which they formed is studied inductively by comparing soils formed from different parent materials but under similar conditions of the other soil forming factors of climate, relief, vegetation, and time. They may be continuous or discontinuous. Continuous lithosequences are illustrated by the spectra of soils formed in loess blankets in Illinois and neighboring states. These loess deposits are thick and relatively coarse-textured at source river valleys, and gradually change to relatively thin, finer-textured layers far eastward from the source areas. Soils formed have a systematic change of several soil properties downwind, including increased clay content in B horizons and exchange acidity (Birkeland 1999).

Discontinuous lithosequences of soils are found on glacial tills of distinctly different lithologies, the composition of which depends on the bedrock sources over which the glaciers passed. Cline (1953) reported major differences in soil properties related to differences in composition and physical properties of such glacial tills (Figure 3.3).

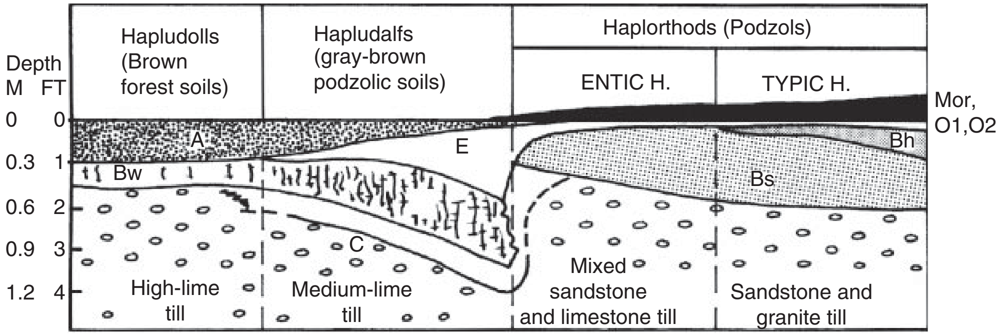

Figure 3.3. Mollisol–Alfisol–Spodosol (Brown Forest–Podzol) lithosequence in New York State. (After Cline 1953)

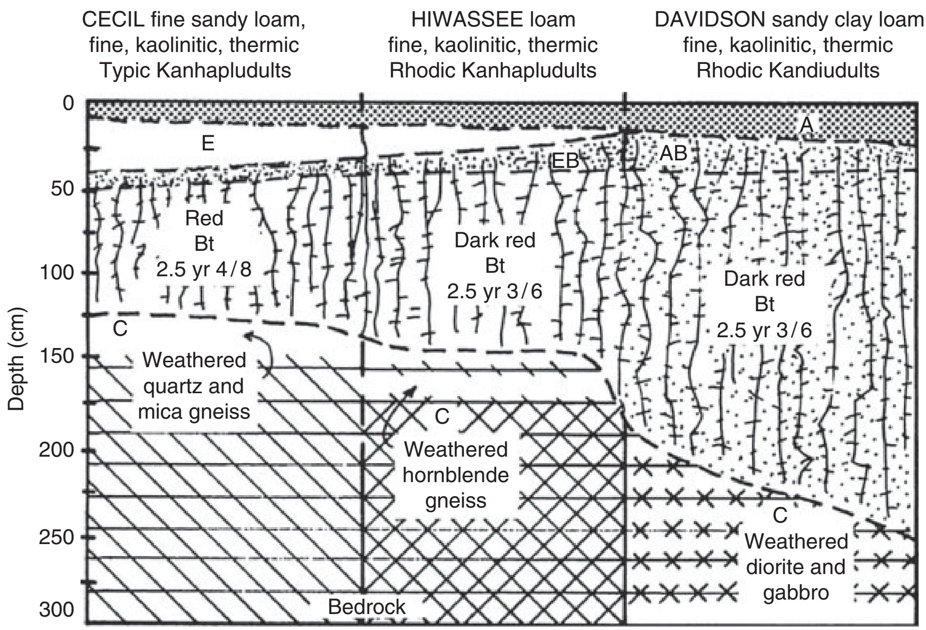

Figure 3.4. Lithosequence of the Piedmont of North Carolina (based on observations by S. W. Buol and R. J. McCracken).

Another classic discontinuous lithosequence consists of soils formed from saprolites derived from contrasting felsic and mafic igneous rocks under similar climatic and vegetative conditions in the Piedmont region of the southeastern United States (Figure 3.4). The Cecil soils formed from felsic, siliceous mica gneiss, or granite gneiss and, as a result, have quartzose sandy A and E horizons. The Davidson soils, derived from diorite or gabbro mafic rocks, differs from the Cecil soils in having a more clayey A horizon, a darker red B horizon with higher iron oxide content, higher pH values, and a lower exchangeable aluminum content. The properties of the Hiwassee soils formed in hornblende gneiss are intermediate between those of the other two soils.

Even minor variations in the amount of certain primary minerals in otherwise similar lithologies can have important influences on soil properties. In North Carolina, soils derived from schist with 7 percent almandine (iron-rich garnet) and 22 percent biotite were redder (5YR hues) and had higher clay contents (31%) than soils formed from gneiss with no almandine and 6% biotite (7.5YR hues, 20% clay) (Graham et al. 1989a; Graham and Buol 1990). Rapid release of iron from the easily weathered almandine produced red-pigmenting hematite and the biotite weathered readily to clay minerals (kaolin and vermiculite). Similarly, tonalite-derived soils in southern California are redder, contain more clay, and are more deeply weathered when the accessory mafic mineral content is higher (5% hornblende, 20% biotite compared to <1% hornblende, 10% biotite) (Graham et al. 1997). In the extreme geochemical environment of ultramafic landscapes, a minor calcium-bearing mineral component can favorably alter the soil fertility, ameliorating the serpentine effect on vegetation and thereby increasing the soil organic carbon content (Lee et al. 2001).

Climate

Climate is an average of extremes, means, and frequency of weather events at a given location. As a factor in soil genesis, climate provides two of the most significant components affecting the processes of soil formation: water and temperature (energy). Although average climatic conditions have long been known to be related to soil properties, extreme weather conditions in many cases may play critical roles in shaping soil properties.

On a global scale, climatic and soil gradients appear to be strikingly congruent, judging by similarity of patterns on world soil maps and climatic maps. Many published works following the publications of Dokuchaev (1883) and Hilgard (1892) give extensive observations on coincidence of climatic and soil zones.

Although climatic patterns relate well to general patterns of soil properties, many contrasting soil properties exist as influenced by parent material and relief (topography) within climatic zones. In this section, both the macro scale relationships and some of the more localized (micro scale) influences of climate on soil formation are examined.

Precipitation–Soil Property Relationships. The role of water in the erosion and deposition of soil material at the soil surface is easily observed. As precipitation increases vegetative biomass usually increases (Jenny 1980), with varying effect on soil properties (Figure 3.5). Water dissolves and transports materials, facilitates growth of plants and other organisms that contribute organic matter to surface horizons. Parallel increase in biomass may not follow unless supplies of plant nutrients are adequate, offsetting higher rates of productivity.

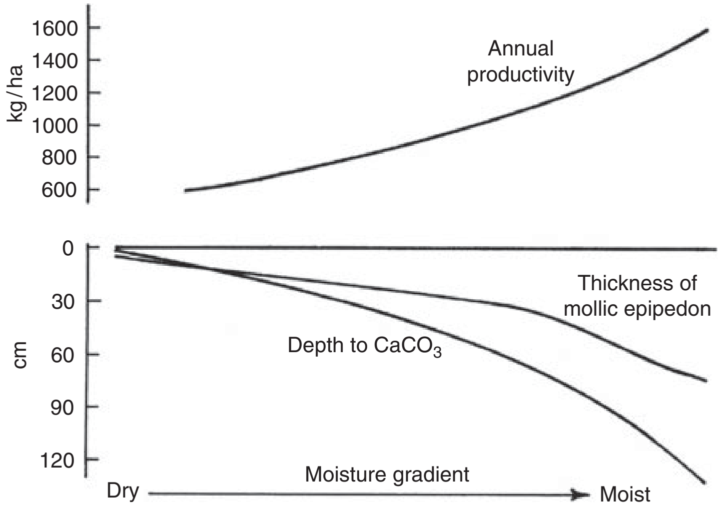

Figure 3.5. The thickness of the mollic epipedon, which is highly correlated with productivity of range vegetation (grass, forbs), proxied for accurate seasonal climatic data from remote desert basin and mountain range sites in Montana (where most climatic parameters had to be estimated), as summarized graphically. Climate becomes both more moist and cooler (*left to right*). (Adapted from Munn et al. 1978, © Soil Science Society of America, with permission)

Monitoring amount and movement of water in particular soil bodies is critical to understanding cause and effect (Bouma 1983; Reddy et al. 2000; Richardson and Vepraskas 2001).

Average annual precipitation on earth has been reported to vary between 1.27 mm (0.05 in.) at Iquique, Chile, South America, and 11,689 mm (460.20 in.) at Waialeale, Hawaii (Reed 1941). Little if any data exist from soils formed under these extreme conditions, but much work has been carried out in intermediate climates to show how soil properties relate to the total amount of rainfall. In summary of that work, Jenny (1941, 1980) reported the following relationships where carbonate was present in the parent material in temperate latitude regions having 380–890 mm (15–35 in.) of annual rainfall:

1. The hydrogen ion concentration in the soil increases (pH decreases) with increasing rainfall.
2. The depth to carbonates in the soil increases as rainfall increases.
3. The nitrogen content of the soil increases with increasing rainfall.
4. The clay content in the solum increases as rainfall increases.

We must be cautious about accepting some of these relationships because of the influence of parent material on clay content. Also, in some very arid areas, parent materials and atmospheric deposition are carbonate free thus the depth to $CaCO_3$ relationship does not exist (Paredes and Buol 1981; Graham and Franco-Vizcaíno 1992).

Average rainfall values have only a regional relationship to soil formation. Within a give climatic region, topographic position greatly alters the proportion of the rainfall that enters the soil or conversely flows over the soil surface at any given site. It is also important to recognize that only a portion of the total rainfall infiltrating the soil surface moves through the entire soil profile. During most rainfall events, infiltrating water only percolates to a modest depth from where it is removed by evaporation or more commonly by transpiration when plants are actively growing. These two forms of water loss from soil are often considered collectively as evapotranspiration. Infiltrated water that only penetrates to a modest depth before it is lost by evapotranspiration tends to remove clay, organic matter, and or soluble salts from surface horizons and deposit them in B horizons. When infiltrated water continues to move below a depth reached by actively growing plants, dissolved colloidal material is leached entirely from the soil. Due to the varying conditions of duration and intensity of rainfall events and seasonal evapotranspiration variations, both scenarios of water movement in the soil are likely to take place in almost every soil except perhaps in the most arid climates.

Several methods can be used for measuring evapotranspiration at a specific site. Not enough data are available from these direct measurements to use in studying regional relationships of precipitation-evapotranspiration (water balance) calculations to soil genesis. However, formulae are available to calculate potential evapotranspiration values from various climatic parameters (Palmer and Havens 1958; Hargreaves and Samani 1986). Although more recent methods are more accurate than ever before (McDonald et al. 1996), the limitation on amount of data available hinders their routine use in studies relating soil development to climatic parameters.

Precipitation is not pure water. The kind and amount of ions reaching the surface of the soil each year vary with amount of precipitation and with location. Some work, utilizing only about 35 weather stations in the United States, plotted by Wolaver and Leith (1972), shows some patterns that should be considered in studies of soil genesis. For example, the central part of the United States—Arkansas to southern Michigan and Wisconsin—receives from 2,000 to 3,400 mg m^{-2} yr^{-1} of Ca^{2+}, whereas eastern Montana receives less than 250 mg m^{-2} yr^{-1} of Ca^{2+}. Potassium and sodium additions appear higher on the coastal areas both east and west, reaching nearly 350 mg m^{-2} yr^{-1} K^+, and 4,000 mg m^{-2} yr^{-1} Na^+ in the coastal areas of Washington and Oregon. Interior parts of the country, Nebraska and Kansas, receive less than 125 mg m^{-2} yr^{-1} K^+, and less than 250 mg m^{-2} yr^{-1} Na^+. Dust deposits of 4.8, 2.8, 1.1, and 0.6 kg ha^{-1} yr^{-1} water-soluble calcium, potassium, magnesium, and sodium, respectively, have been measured in Niger, West Africa (Drees et al. 1993). Locally, most likely due to human activity around cities, yearly additions of SO_4^{2-} and NO_3^- may be higher than in more rural areas. Sulfur deficiency is common throughout tropical latitudes (Sanchez 1976). It is commonly believed that lesser amounts of industrial smoke in tropical areas may account for the widespread deficiencies.

Temperature–Soil Property Relationships. Temperature as a component of the climate influences many reactions involved in the processes of forming soil. With increasing temperature, soil colors tend to become less gray and more reddish. Nitrogen and organic carbon contents decrease as the temperature increases (Jenny 1941, 1980).

Except for additions of heat energy transmitted to the soil from subterranean sources, which may be of local importance in geyser basins, metabolic heat from organisms in the soil, and heat from exothermic chemical reactions in the soil, all of which are considered minor, the predominant source of energy is the sun. Solar energy arriving at the soil surface varies widely, depending on several factors such as day length and cloud cover. Solar radiation increases with elevation. The rate of radiation increase is most rapid in the lower dust-filled layers of the air (Geiger 1957).

Energy transfer involved in heating and cooling the soil takes place at the upper boundary of the soil. Heat absorbed during the day is lost during the night. This loss takes place by reradiation (radiation from the soil into space). In general, the rate of reradiation for soil surfaces can be calculated as if the soil were a black body. According to the Stefan-Boltzmann equation, the rate of reradiation in calories $cm^{-2}\ min^{-1}$ (S) is: $S = S(T)^{4}$ where S has a constant value of 8.25×10^{-11} and T is temperature in absolute or Kelvin degrees. The magnitude of the temperature change in the soil depends not only on the features affecting radiation, but also on the thermal properties of the soil and amount of turbulence created by wind. A schematic graph of the diurnal temperature change in a soil profile is given in Figure 3.6 and demonstrates that the greatest temperature changes take place near the surface of the soil, and that diurnal temperature changes are absent below about 50 cm (20 in.). Water vapor and other gases in the soil move from warmer to cooler sites. Seasonal and daily dynamics of soil temperature cause changes in the rate of CO_2 and other gas emissions from the soil (Smagin 2006).

Air temperature is the principal component in the calculation of potential evapotranspiration and thus has a large measure of control on the depth of water movement within soil. When soil cools to the point at which water solidifies, all of the chemical reactions involving water cease, but physical breakdown of larger particles by frost action may occur. In dry valleys of Antarctica (Pastor and Bockheim 1980), however, this does not produce appreciable amounts of clay in soils that are several million years old but lack Bt horizons.

Temperature exerts a marked influence on the type and quantity of vegetation present in an area. Thus, to a large extent, it controls the amount and kind of organic matter produced. Temperatures also determine the rate at which microorganisms in and on the soil surface decompose organic material. Each soil has a characteristic organic carbon-retention capacity that can be considered a steady state reaction wherein input rate of organic carbon equals decomposition rate. Several factors such as composition of the original organic material and subsequent decomposition products including organic-mineral complexes influence organic carbon retention, that is, organic carbon content, and maximum soil temperature appears to be a critical and perhaps controlling factor (Chenu 2006).

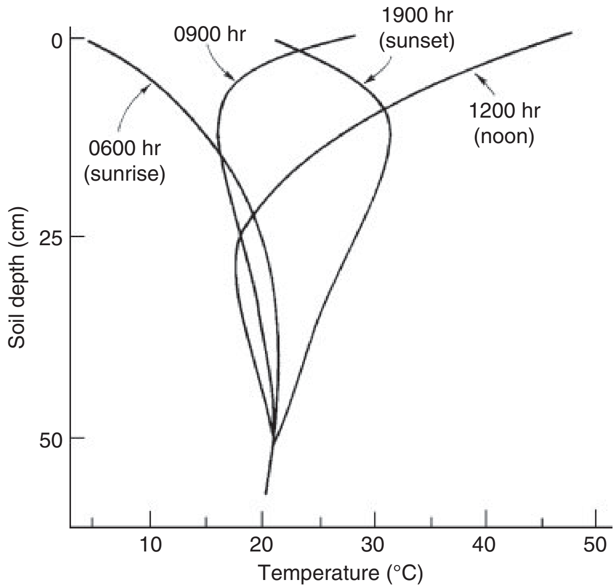

Figure 3.6. Schematic graph of diurnal soil temperature variation with depth.

The influence of temperature on chemical processes in soil can readily be seen by considering Van't Hoff's temperature rule (Van't Hoff 1884), which can be stated: "For every 10°C rise in temperature the speed of a chemical reaction increases by a factor of two to three." Fang and Moncrieff (2001) determined that respiration rate in soil cores increased exponentially with temperature. This relationship is seen as responsible for the higher soil organic carbon contents observed in colder mean annual temperatures in the United States (Jenny 1941). However, when soils from Colombia, South America, were studied their soil organic carbon contents were higher than predicted by temperate latitude mean annual temperatures (Jenny et al. 1948).

When comparing soil organic carbon contents to 30 cm depth among the various soil temperature regimes, it was found that soils with nearly constant soil temperature, that is, iso-soil temperature regimes of the tropics with soils with comparable mean annual temperature regimes in temperate areas, had significantly higher organic carbon contents (Figure 3.7).

In part this can be because summer temperatures in temperate latitudes, non-iso soil temperature regimes, are much higher than the mean annual temperature, whereas in the tropics, there is little seasonal temperature variation. Thus, it appears that the seasonal high summer temperatures and the more rapid rate of organic decomposition they induce are significant factors responsible for determining soil organic carbon content although increased inputs of organic carbon from longer growing seasons in tropical conditions where moisture supplies are not limiting may also be a factor.

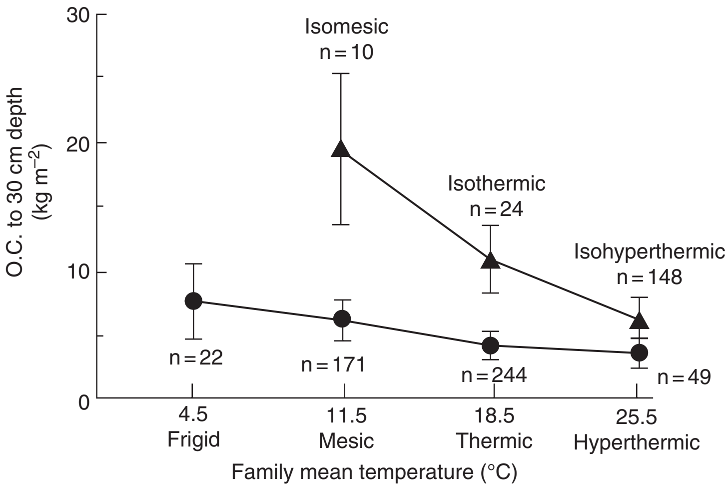

Figure 3.7. Organic carbon content in top 30 cm of soils in soil temperature families. (Organic carbon values plotted to midpoint of soil temperature families.) Source: Buol et al. 1990.

Solar radiation available for heating the soil changes markedly with the season of the year in temperate latitudes and is greatly reduced by cloud cover. Once the incoming solar radiation has penetrated the atmosphere, it has to be absorbed before it is changed into heat. The absorption of the solar radiation at the soil surface is affected by variables such as soil color, orientation of the surface with respect to the incoming solar radiation, and vegetative cover. In general, the darker the color, the more radiation is absorbed and the lower the albedo. The nearer the soil surfaces approach an angle normal to the direct rays of the sun, the greater the absorption. The effect of vegetative cover on absorption varies with density, height, and color of the vegetation. In many forested areas, the vegetation almost completely shades the soil surface. In the case of deciduous trees in the temperate latitudes, however, the soil surface may have little shade in the early spring and almost complete shade in the summer. When forest vegetation is removed and soils are cultivated, their surfaces are not shaded for longer periods of time. Removal of forest vegetation in Ghana resulted in daily maximum surface and near surface soil temperature increases with a corresponding decrease in soil organic carbon content (Cunningham 1963). In Nigeria, Lal and Cummings (1979) measured maximum daily soil temperatures at a depth of 1 cm to be 20–25°C higher in sites where the vegetation was cleared by bulldozer and slash and burn techniques than in adjacent sites with natural forest vegetation.

In part decreases in soil organic carbon contents observed when the shade of native vegetation is removed and soils cultivated, thus exposed to direct sunshine for some periods of time can be attributed to higher soil temperatures and increased rates of organic carbon decomposition.

The amount of water in a soil strongly influences the maximum daily temperature of near surface layers of a soil. The specific heat capacity of water is 1 calorie gram^{-1} (cal g^{-1}), that is, 1 calorie of heat raises 1 gram of water by 1°C. The specific heat capacity of most oven dry soil is 0.2 cal gm^{-1}. If an oven dry surface soil of 1 g cm^{-3} bulk density is heated by 1 calorie of energy the temperature rise will be 5°C. If the same volume of soil was saturated with water that would occupy the 50 percent void space in most soil material, that is, 0.5 g water with no volume change, the specific heat capacity of the saturated soils would be 0.7 cal cm^{-3} and 1 cal of energy would raise the temperature only 1.43°C. Thus, we consistently find that soils that are frequently saturated (aquic soil moisture regimes) retain greater quantities of soil organic carbon than unsaturated soils. Similar reasoning can account for higher soil organic carbon contents in clay-textured soil material that often retains approximately 20 percent water at permanent wilting moisture content than in adjacent sandy-texture soil that only retains less than 10 percent water at permanent wilting.

Thornthwaite (1948) developed an empirical formula to express the relationship between potential evapotranspiration and temperature. Palmer and Havens (1958) facilitated the use of the Thornthwaite formula by developing nomographs. The Soil Conservation Service (USDA 1960), seeing the usefulness of the Palmer–Havens nomographs, reproduced them on a larger scale. The Thornthwaite technique or a number of other techniques can be used for arriving at a potential evapotranspiration value (Hargreaves 1974).

Monthly normal precipitation and calculated potential evapotranspiration values are used in *Soil Taxonomy* to calculate the soil moisture regimes. Such calculations only apply to soils in the area when the following assumptions are made: (1) there are no additions or losses of water by runoff; (2) vegetation is growing on the site and actively transpiring; and (3) dew formation is negligible.

With these assumptions a simple water balance for a soil illustrated in Figure 3.8 where annual precipitation $P = 125$ cm (49 in.) with annual potential evapotranspiration $e = 94$ cm (37 in.) indicating that 30 cm (12 in.) of surplus water (s) percolates below the lower boundary of the soil, that is, $p - e = s$.

A water-storage capacity of 22 cm (8.7 in.) is used in the illustration to better illustrate how water balance calculations can be used to evaluate processes of soil formation and not the 5 cm (2 in.) used in soil moisture regime calculations of *Soil Taxonomy*. In this example, the following observations can be made:

1. Average monthly rainfall (the dashed line) is greatest in March and December.
2. Potential evapotranspiration is greatest in July and lowest in January.
3. Precipitation exceeds potential evapotranspiration from November through May. (Note: P and e for each month are plotted at the end of the month.)
4. Potential evapotranspiration exceeds rainfall June through October.

The seasonal moisture dynamics that influence soil-forming processes in the soil can be interpreted as follows. During the summer, all of the rainfall and the water stored

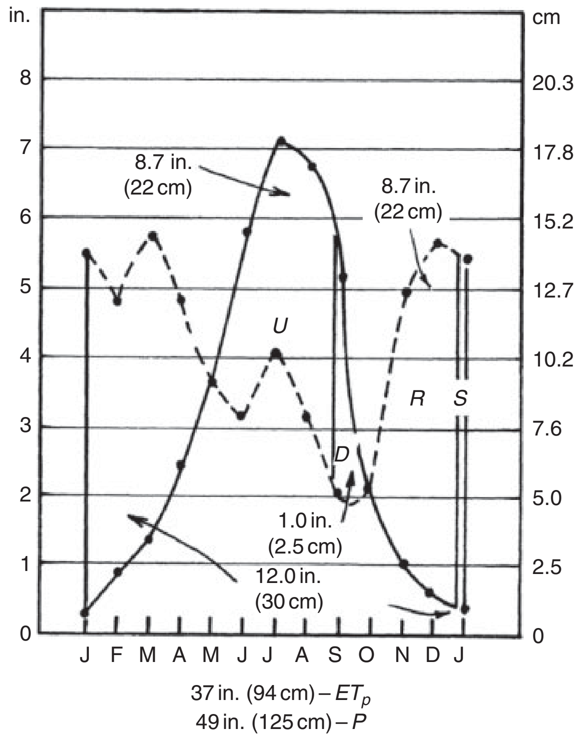

Figure 3.8. Average yearly water balance at Tallulah, Louisiana, for a soil storing 22 cm (8.7 in.) of water. Note that this is a total water storage value, not a moisture control section storage value.

in the root zone (22 cm) are utilized (*U*) via evapotranspiration. During this time, water from each rainfall event percolates only to a shallow depth before it is transpired, depositing any suspended or dissolved material it may contain. In September there is a short period when there is a deficit (*D*) of rainfall and water storage compared to evapotranspiration, and plants can be expected to wilt to conserve limited water. In October rainfall exceeds potential evapotranspiration, and available water supplies are recharged (*R*). From December to May, with the storage capacity of the root zone full, any rainfall is surplus (*S*) and water, containing any dissolved and suspended material that it may contain leaches below the rooting depth presumably reaching the groundwater.

Another quantity, calculated actual evapotranspiration (*Ea*), can be determined from the water balance calculations. To calculate *Ea,* a water-holding capacity for root zone of specific plants needs to be determined or estimated. Again using Figure 3.8, in which the water-holding capacity is 22 cm (8.7 in.), the calculated actual evapotranspiration is the sum of the potential evapotranspiration values (*e*) for the months during which precipitation (*P*) exceeds potential evapotranspiration (*e*), that is, November through May plus the potential evapotranspiration for that period during which stored water is utilized, that is, June, July, and part of August, plus the precipitation (*P*) for the months in which there is a moisture deficit (*D*). Actual

Table 3.5. Summary of microclimatic conditions on north- and south-facing slopes in Michigan during the 1957 growing season

Measurement	Slope Facing	Mar. 12– Apr. 14	Apr. 15– June 1	June 2– July 15	July 16– Aug. 23	Aug. 24– Sept. 16	Sept. 17– Oct. 16	Oct. 17– Nov. 9
Rel. light int. (% in open)	North	34.2	21.1	2.5	1.8	1.8	2.4	20.5
	South	42.4	30.7	7.0	3.5	3.8	5.5	36.8
Max. air temp. (°F) 50 cm	North	48.5	72.9	79.6	85.8	75.8	70.4	61.3
above soil	South	52.5	76.7	87.7	95.1	80.3	76.9	66.7
Min. air temp. (°F) 50 cm	North	25.7	46.5	58.1	57.1	51.8	39.4	29.4
above soil	South	27.1	47.7	58.2	57.5	52.2	39.8	29.6
Evaporation cc/wk	North	N.R.	99.4	75.1	84.8	61.4	59.8	N.R.
	South	N.R.	144.6	118.1	119.3	95.8	91.4	N.R.
Soil temp. (°F) at 2 cm	North	34.8	57.6	66.4	70.8	65.6	55.8	47.4
	South	41.1	61.1	69.5	76.1	68.2	60.3	51.0
Soil temp. (°F) at 20 cm	North	33.0	50.5	61.8	65.5	62.4	55.0	48.4
	South	38.8	55.6	64.9	70.6	65.6	59.6	50.7
Percent moisture at 2 cm	North	30.2	27.4	19.4	13.4	11.6	9.2	16.2
(% by wt.)	South	17.5	19.4	15.4	8.8	10.2	6.8	14.4

Source: Cooper 1960.
Note: N.R. = not recorded.

evapotranspiration values are sometimes used to evaluate the water available for plant growth in moisture-limited soils.

Arkley (1967) used water balance calculations to characterize several kinds of soil in the United States. He reasoned that the degree of leaching would be best characterized as the annual sum of monthly $P - e$ values from those months when $P > e$ or the amount of precipitation in the wettest month, whichever was greatest. He called this value the leaching index (Li). Arkley reasoned that a greater amount of actual evapotranspiration would be an index of several soil properties. In particular, he reasoned that with more *Ea,* more energy would result in greater organic matter production and weathering. When climatic data associated with great soil groups, as mapped on available small-scale maps, were plotted on charts with Li as one axis and calculated actual evapotranspiration (*Ea*) as the other axis, clear parameters could be defined for each great soil group. Considerable overlapping of great soil groups was observed, but much of this could be expected because of differences in water-holding capacity (a figure standardized at 6 in. in Arkley's calculations), the general nature of the maps available, and the lack of a quantitative range of characteristics used in defining the great soil groups of the 1938 Soil Classification System (Baldwin et al. 1938).

Water balance calculations have also been used to calculate the age of soils containing calcium carbonate (Arkley 1963; Jenny 1980, p. 189; McDonald et al. 1996). By determining the water-holding capacity of the various soil horizons, the carbonate solubility, and carbonate content of the parent material, and by calculating the amount of water passing a given depth in the soil from water balance data, soil ages were obtained that closely correlated with ages indicated from geologic data.

Cool winter rains in California seem to be more effective at solubilizing and leaching calcium carbonate than warm summer rains of the central Great Plains of North America (Jenny 1980), as would be expected, considering greater solubility of CO_2 in water at lower temperatures. On the other hand, desilication in soils is mainly dependent on the amount of water available for leaching, because only a slight increase in solubility of silicon results from rise in temperature (Yaalon 1983).

Microclimate and Soil Genesis. Microclimate can be thought of as that climate in the first few meters above the soil surface. It varies markedly with relief features, soil color, and vegetation.

In the Northern Hemisphere, south-facing slopes tend to be warmer and droughtier than north-facing slopes. In a comparison of north- and south-facing slopes in Michigan, detailed comparisons were made during the growing season in 1957 (Cooper 1960).

On the slopes where the data in Table 3.5 were obtained, the soils on the south-facing slope were found to have lighter brown A horizons and redder B horizons than those on the north-facing slope. Soils on the south-facing slope contained more clay in the B horizon (average 12.6 percent) than did the north-facing soils (average 7.7 percent). Also, the north-facing soils had both deeper solums and A horizons, 91–103 cm and 65–66 cm, respectively, compared with south-facing soils, 64–85 cm and 40–64 cm, respectively.

Shul'gin (1957) reported that similar north–south relationships were found in Russia and that soils on west-facing slopes were warmer than those on east-facing slopes. He attributed this to a greater expenditure of solar insolation for evaporation of dew on east-facing slopes than on west-facing slopes, which were drier during the afternoon when they received more direct insolation. He reported that a field sloping 1 degree toward the south was as warm as a level field 100 km farther south. The magnitude of the temperature–slope aspect decreases in more equatorial areas.

Cold-air ponding or cold-air drainage into depressions in the landscape alters the climate of the local sites. Cold air, because of its greater density, tends to move down slopes during the night and to collect in valley bottoms. It is in these depressions that coldest air temperatures have been recorded. Frosts are more likely to occur in these areas.

The properties of the soil in some measure tend to control the climate in which they form. Light or whitish-colored soil surfaces tend to reflect more radiation than darker-colored soils. When incoming solar radiation is reflected, there is less net radiation to be absorbed and heat the soil. Snow is especially effective in reflecting the incoming radiation.

Climosequences. The various soil properties discussed in relation to moisture and temperature regimes form a continuous pattern over the landscape.

Climosequences are most often studied in mountainous areas where the temperatures decrease at a rate of about 6°C per each additional 1,000 meters of elevation (Petterssen 1941), and rainfall usually increases concurrently. Although the rate of change varies, the following observations have been reported (Fridland 1959; Martin and Fletcher 1943; Vologuev 1964; Whittaker et al. 1968; Alexander et al. 1993; Dahlgren et al. 1997): organic matter content, nitrogen content, and C/N ratio increase with increased elevation, whereas pH value and calcium, magnesium, and potassium contents decrease with increased elevation.

Climosequences are dramatic when compressed into the space of a few kilometers on a mountainside (Jenny 1980) but are best observed over longer transects where the changes in soil characteristics occur on more gradual slope gradients.

Climate is not constant with time. Although meteorological measurements do not reach into the past for any great length of time, there is much geologic and botanical evidence to show that climate has changed through time (Ruddiman 2001). There is clear evidence that climatic change has been important within the history of many of the soils we currently observe on the earth's surface. Soil changes in response to changes in its environment so that only irreversible characteristics are likely to remain (Nikiforoff 1953).

Climate can be considered as an average of weather events. We have all observed that one year is not exactly like any other year. Throughout the history of soil science, attempts have been made to establish a quantitative measure of the relationship of a

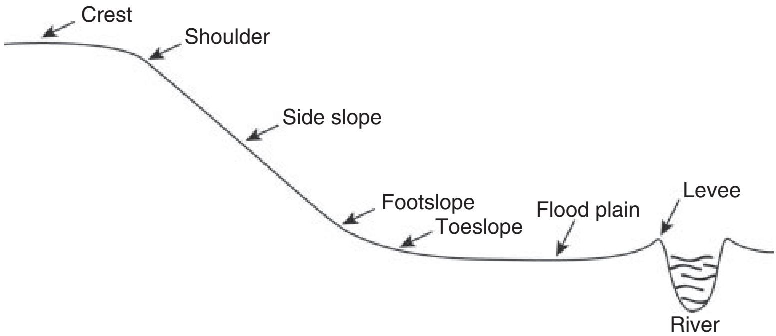

Figure 3.9. Cross section diagram of landscape positions.

certain soil property to various climatic elements. This work has led to a better understanding of soil–climate relationships discussed in this chapter. It should not be anticipated, however, that these relationships can ever be exact when only climatic averages are used. The event in the weather that produces a given soil property may be caused by certain infrequent climatic circumstances. For example, in an arid region, a 1-year-in-100-rainfall event may produce enough infiltrating water to leach soluble salts from the soil. If the soluble mineral content is small, the soluble salt content of the soil may remain low for several years of more normal precipitation.

Relief (Topography)

Relief (*r*) is the physical configuration of the land surface with respect to relative elevation and slope. Attributes of relief include elevation above a datum (usually meters above sea level), slope and natural soil drainage condition, aspect, and slope configurations of convexity and concavity (Moore et al. 1993). Ruhe (1969a), Jenny (1980), Birkeland (1999), and Wysocki et al. (2000) have treated topographic relationships of soil extensively.

A cross section overview of landscape positions is illustrated in Figure 3.9. The highest elevation or summit position is represented with a rather gentle slope. At the edge of the summit is a narrow zone of steep slopes called the shoulder position below which is a linear area of somewhat lesser slope called the back slope. At the bottom of the back slope as the angle of the slope decreases, there is a foot slope position followed by a toe slope position of even lesser degree of slope. Below the toe slope is a nearly level floodplain area that has as on its boarder with a river a slightly elevated rim called the natural levee.

Each component of the landscape has a relationship to the surface flow and erosive processes of water. The summit position, with gentle slopes, normally infiltrates most precipitation and contributes runoff only during intense rainfall events. The shoulder position thus receives little sediment from the summit but being at the top of the more sloping back slope experiences more rapid loss of material from erosion. The back slope while subject to erosion loss also receives some sediment from the shoulder during brief rainfall events. The back slope position is also subject to the creep that further enhances the slope angle at the shoulder position. At the base of the back slope, as the slope angle decreases in the foot slope position, the velocity of the runoff decreases as it encounters vegetation and sediment tends to be deposited. The foot slope position is also the recipient of creep material from the back slope. As the slope angle further decreases below the foot slope, a toe slope position sometimes barely distinguishable from the foot slope, receives sediment from above and is rarely eroded. On the floodplain, most sediment comes as the river overflows its channel during intense rain events. As water escapes the river channel and spreads over the floodplain, velocity decreases and suspended sediment is deposited. The first sediment deposited is coarse sand-sized particles resulting in a sandy textured, slightly elevated levee adjacent to the river channel.

Among the positions illustrated in Figure 3.9, a rather predicable pattern of soil characteristics is developed. On the summit soils tend to have fully developed properties characteristic of the climatic, vegetative, and parent material conditions of the area. The shoulder positions have the thinnest soil profiles (Nikiforoff 1949; Graham et al. 1990). The back slopes develop soil profiles of intermediate thickness between soils on the shoulder and soils on the summit. On foot slope and toe slope positions, significantly thicker soil profiles, especially A horizons, are usually present (Riecken and Poetsch 1960). Soils on the floodplain are usually thick and have stratified layers with abrupt boundaries, some of which have high organic carbon content. A gradient of sandier to more clayey particle size is usually present from near the river toward the boundary with the toe slope.

The degree to which hill slope position affects soil properties varies with vegetative cover, texture, aggregation, porosity, and shrink–swell potential of the soils and to a large extent on the intensity of the rainfall events. Where rainfall events are seldom intense, the effects of hill slope position are less apparent than in areas where intense thunderstorms are more frequent. Thickness of solum development is inversely related to angle of slope. Norton and Smith (1930) reported that loess-derived soils had solum depths from 20 to 35 cm on slopes of 14 and 8 percent, respectively, to 63 cm on 1 percent slopes. In western Iowa, soil organic carbon contents were found to be lowest on ridge crests (summit position) and highest at low-lying slope positions (Aandahl 1949).

Hill slope position is often reflected in growth rate of trees. Wilde (1958) summarized a general observation of greater tree growth on lower side slopes and foot slopes attributed to enrichment of fertility via surface deposition and run-on water. As a result he reported that foresters often say, "Trees level the relief with their crowns" (p. 248).

Conflicting relationships of hill slope position and specific profile features such as E horizon thickness have been reported. Låg (1951) found thicker E horizons in Spodosols (Podzols) in depressions than in similar but drier soils on adjacent uplands. The opposite was reported for Ultisols on the North Carolina coastal plain, where the

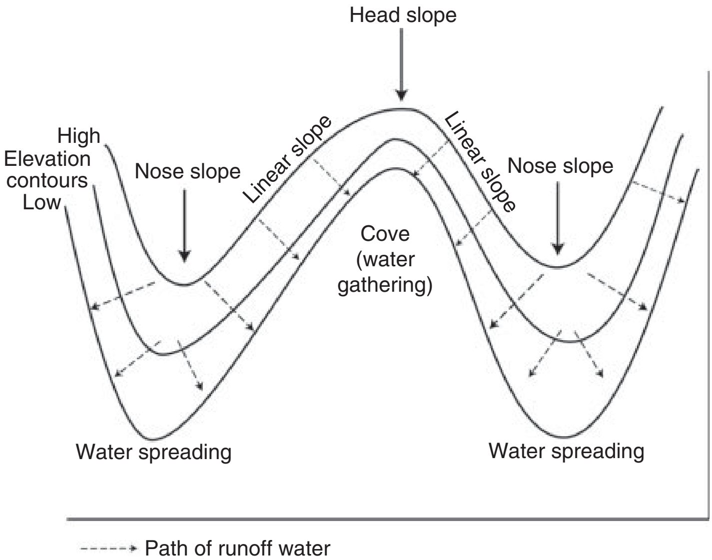

Figure 3.10. Diagram of hill slope positions.

E horizon reached maximum thickness on upland areas rather than in depressions (Daniels et al. 1967).

Slope Configuration and Positions. Slope configuration at a site can be described as concave and convex in both plan and elevation view or convexo–concave with summit-to-toe slope sequences (Figure 3.10). These slope parameters correlate with the disposition of water and soil properties (Graham 2006).

Contrasting conditions between head slopes (water gathering), nose slope (water spreading), and side slopes (water paralleling) are particularly associated with variation of soil properties, including hydrologic ones. Cove positions, water-gathering areas commonly have thicker soils due to sediment deposition and increased moisture. In some cases they are sites for gully erosion if runoff is excessive.

Relief is modified by erosion during soil formation. Stair step configurations sometimes result when erosion-resistant layers in parent material and/or soil horizons are encountered (Figure 3.11).

On small-scale soil maps, variations in soils between head, nose, and side-slope positions are commonly not shown. Omission may be a result of cartographic generalization required by map scale, or alteration of original soil pattern by cultivation and grazing. Research maps (Stone et al. 1985; Daniels et al. 1985) may reflect patterns of local erosion and sedimentation in a variety of landscape positions.

In addition to losses and gains of material by localized erosion and deposition related to slope, there is also creep or solifluction of the entire soil mass downslope. Direct measurement of soil creep, by means of simple yet accurate devices, showed 11.5 mm of movement in a 2-year period at a depth of 15 cm, and 1.2 mm of movement

Figure 3.11. Downwasting landscape where hilltops are protected by plinthite layers in the Llanos of Colombia, South America.

at a depth of 105 cm, on a 4 percent slope in Iowa (Troeh 1975). In New South Wales, Australia, Heirnsath et al. (2002), using optically stimulated luminescence of quartz grains, reported average slope-parallel (creep) velocity of approximately 3.5 cm yr^{-1} on slope gradients of 15–45 percent. Considerable variation in soil properties, saprolite, and colluvium thickness was reported on hill slopes of 10–75 percent in the Blue Ridge Front in North Carolina (Graham et al. 1990).

Soils of Open and Closed Drainage Systems. The diagram in Figure 3.10 relates to open drainage systems noting contrasting soilscape positions in that context. Closed systems, from which particulate sediment does not escape in running water, are in landforms that include sinks in carbonate-rich materials, kettles in glacial drift, inactive volcanic craters, depressions between sand dunes, and oxbow depressions in alluvial deposits. Sodium-rich natric horizons form in and around small seasonally wet depressions within closed drainage systems in North Dakota (Seelig et al. 1990). In semi-arid environments, closed basins often support vernal pools and have clayey, smectite-rich soils with higher levels of manganese oxides than those upslope (Weitkamp et al. 1996). Walker (1966a) has reported in detail soil sequences that evolved in a bog basin in Iowa. Concentric soil patterns develop in and around such features. In depressions, soil microclimate may be hydric to xeric, depending on macroclimatic precipitation, and temperature may be elevated by insolation or depressed by drainage of cold air into the basins. On a small scale, depressions 1- to

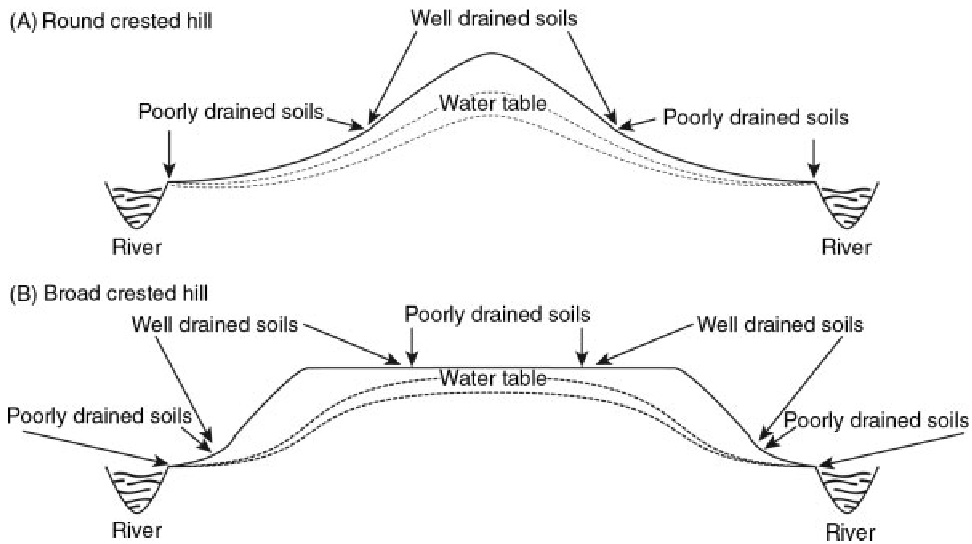

Figure 3.12. Schematic sketches of some soil catenal relationships in (A) a relatively unleached, originally calcareous soilscape with wet, level areas flanking the interfluve, and (B) a highly leached, coastal plain soilscape with wet, level areas on the interfluve (C = small areas of enriched, poorly drained soils).

2-m across are associated with tree-tip mounds, and hollows about 6 m in diameter lie between hummocks called Mima mounds (Jenny 1980).

Relationship of Topography and Water Table: Catenas and Toposequences. In humid climatic conditions, udic and some ustic soil moisture regimes where water tables are near the soil surface, relief soil properties have contrasting relationships depending upon topographic configuration (Figure 3.12). A water table is the upper surface of saturated conditions that seasonally and sporadically, in response to rainfall events, invades soil profiles and in some areas rises to the soil surface initiating chemically reducing conditions in the soil. Soils so affected are classified as having aquic conditions and/or an aquic moisture regime. Detailed soil surveys use the term "poorly drained" and many such soils are now popularly termed "hydric" soils in definitions of "wetlands."

In Figure 3.12A areas where the topography consists of rounded hills and valleys, the relationship of water table depth is depicted with soils commonly named as "well-drained," "moderately well drained", "somewhat poorly drained," and "poorly drained" forming a sequence (toposequence) as the water table becomes closer to the soil surface in the lowest elevations. This sequence of soils is commonly referred to as a *catena* (Milne 1935). Bushnell (1942) defined a catena as a toposequence formed on a single parent material, although allowing the lowest-lying member to be formed from organic materials entirely, whereas the more elevated members were mineral soils.

However, in humid areas with nearly level topography, the arrangement of catenas with respect to topography becomes more complicated and in some parts of the terrain are reversed (Figure 3.12B). In response to the slow movement of groundwater, broad, nearly level areas, often the highest elevations in the area, experience a mounding of the water table and the soils are *poorly drained*. A sequence of drainage conditions radiates outward and *well-drained* soils are present on and near the shoulder of the slope, and from there to the lowest points on the landscape, a sequence of *well* to *poorly drained* soils is present.

Organisms

Flora and fauna, including the human species, constitute both an internal and external biotic factor (ϕ), in soil formation. The current vegetation and animal life at a site are often the survivors of a succession of organisms that occupied the site and influenced soil formation. Multitudes of microorganisms inhabit the soil. Distribution of species and communities of organisms are related to climatic environments, and thus direct comparisons are confined geographically and are not possible among all soils. Of all the soil-forming factors, organisms are the least susceptible to numerical expression. Perhaps numerical codes may be developed some day to quantify the effect particular organisms or their communities have on soil (Ugolini and Edmonds 1983).

Soils and organisms interact; therefore, a soil and its biota may be said to evolve together (Jenny 1980). A soil is not only a biotic environment of plants, it is also a participant in plant growth and succession. For example, on marine terraces beside Glacier Bay, Alaska, growth of western hemlock forest and formation of Spodosols have been simultaneous (Ugolini and Mann 1979). Placic horizons have formed in maturing soils and have given rise to perched water tables that have allowed sphagnum moss to invade and replace the forest. A bog soil (Fibrist) has buried the Spodosols in places, as a climax in the synchronous biotic and pedologic successions. The role of the multilayered biotic community that surmounts and permeates the horizons of a soil body is significant in soil genesis.

Vysotskii (1957) spoke of the soil as the arena of life and water as the lifeblood of the soil. "Life is the process of maintaining bodies of protein," a process in which the soil is intimately involved. Polynov (1951) noted that the relative mobility from land to sea of the four elements Na > K > Ca > Mg has been changed by soil and biota to Ca > Na > Mg > K. Since life first appeared on this planet, the biosphere has been accumulating oxygen in the atmosphere, calcium in fossiliferous limestone, and carbon in fossil fuels. The evolutionary succession of conifers, angiosperms, and grasses has meant increasing efficiency in biocycling phosphorus and in protecting the soil against erosion.

Living today are about 450,000 kinds of plants, 1.5 million kinds of animals (of which insects are the most numerous), combined in an infinite number of biotic communities. Land management by people greatly affects the interactions between biota and soils.

Table 3.6. Average carbon to nutrient element ratios (g C/g nutrient) and pH values of forest floor (O horizons) for seven tree species

Species	C/N	C/P	C/Ca	C/K	C/Mg	pH CaC_{12}
Lodgepole pine	35.2	674	264	805	753	3.48
Sitka spruce	28.7	530	94	533	648	3.99
Norway spruce	26.4	462	77	412	480	4.26
Douglas fir	25.6	452	114	462	546	3.98
Grand fir	31.4	434	58	438	482	4.46
Beech	26.8	465	48	337	396	4.63
Oak	27.5	440	55	315	398	3.97

Source: Vesterdal and Raulund-Rasmussen 1998.

Inputs of organic matter with its major components C, O, and H obtained from CO_2 in the air and water are prerequisite to soil formation. Net primary production of organic matter in an ecosystem is about half of gross primary production because of the large amount of material consumed in respiration by plants. Schlesinger (1997) reports figures for net primary production (g C m^{-2} yr^{-1}) of various broad ecosystem types, ranging from 80 in deserts to 1,300 in wetlands, 130 in tundra, 800 in wet, tropical forests, and 650 in temperate forests. Total biomass of flora and fauna during a favorable season is probably 2 to 30 times this amount, the larger factor being applicable to forest ecosystems in which large masses of dead heartwood are usually counted as part of the live stand.

The chemical composition of the organic residue deposited by different types of vegetation is species dependent. Vesterdal and Raulund-Rasmussen (1998) analyzed the forest floor (O horizons) under even-aged, 30-year-old stands of seven species in Denmark. The species studied were lodgepole pine (*Pinus contorta*), Sitka spruce (*Picea sitchensis*), Norway spruce (*Picea abies*), Douglas fir (*Pseudotstuga menziesii*), grand fir (*Abies grandis*), Beech (*Fagus sylvatica*), and oak (*Quercus robur*). The sites were almost level research plots planted on seven contrasting soil conditions ranging from fertile Mollisols and Alfisols to infertile Inceptisols, sand Entisols and Spodosols. Although minor chemical differences in forest floor composition could be correlated to type of soil, the major differences were related to tree species (Table 3.6). The most significant chemical differences were between the lodgepole pine and oak forest. The total amount of carbon in the lodgepole pine forest floor was approximately six times greater than in the oak forest floor. The amount of carbon in the forest floor over the most infertile soils was approximately three times greater than in the most fertile soils. It appears that the extremely high carbon to nutrient element ratios (Table 3.6) limit the decomposition rate of forest litter resulting in greater amounts of nutrient-poor material and lesser amounts of nutrient release to be biocycled and reused by the growing vegetation.

Unicellular and acellular organisms (protistans), chiefly the microflora and microfauna, occur in vast numbers (Figure 3.13) in soils. They capture nitrogen (N_2) from the air, decompose organic debris, produce both growth-promoting and

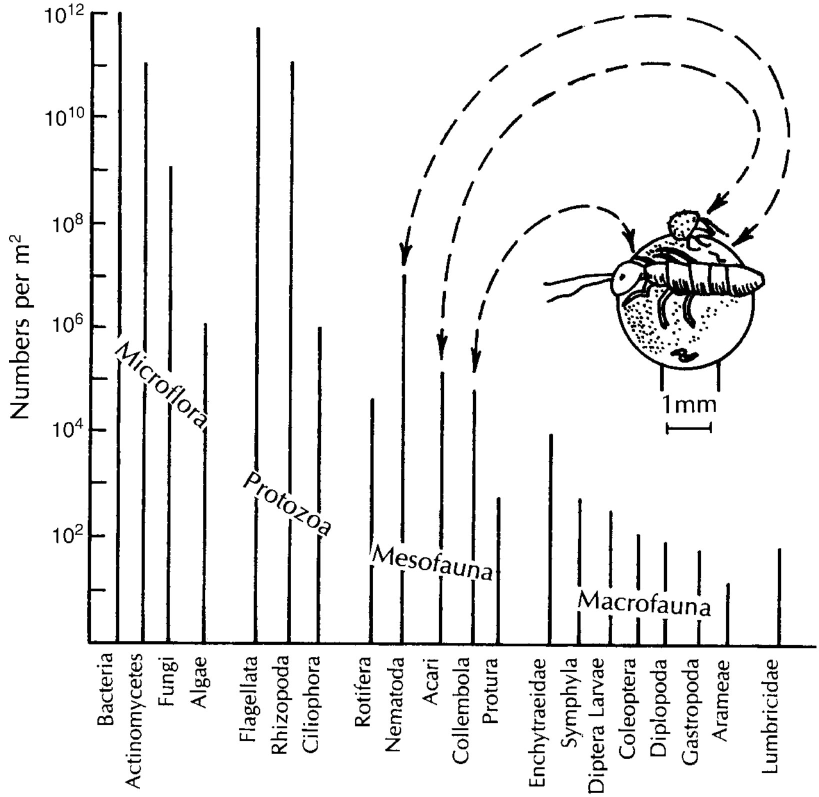

Figure 3.13. Approximate numbers of representative soil organisms per square meter (surface area) of a pedon in a humid temperate zone. Size categories are those of Drift (1951) (see Hole 1981, Figure 3). Three kinds of animals are sketched above (*right*) as if perched on a pinhead.

growth-inhibiting compounds for a variety of organisms, stabilize and mobilize metallic ions and colloids, and serve as a foundation for "food chains" for organisms in soil. If protistans ceased to function, cycles involving nutrition of other organisms and disposal of biotic debris would fail, and the protective, biotic blanket over the fragile "geoderma" would disappear.

The generalization has been made that a gram of soil may contain a million bacteria, a kilometer of fungal hyphae, and thousands of algal and protozoan cells. Yet, there are scantily populated microsites within the same gram of soil, partly as a result of insufficiency of certain growth requirements and partly the effect of diffusion of growth-inhibiting substances. Aerobic and anaerobic conditions may coexist within a few millimeters of each other, depending on size of soil peds and sequence of hydrologic events in a soil horizon. This heterogeneity gives rise to a complex distribution of microbial species within soil.

Biosequences of Soils. Plants, differentiated from animals by the lack of locomotive movement and sensory organs and by the presence of cellulose cell walls, may be

Table 3.7. Some properties of well-drained soils developed under prairie, prairie border, and nearby deciduous forest cover in the North Central Region of the United States

Soil Property	Soil Formed under Prairie	Soil Formed under Prairie Border Conditions	Soil Formed under Deciduous Forest
Clay content (%)			
A horizon	28	23	21
B horizon	34	36	36
Ratio, B: A	1.21	1.57	1.71
Total nitrogen (%)			
A1 (0–7 cm)	0.35	0.35	0.35
E (20–30 cm)	0.21	0.11	0.08
Bt (55–65 cm)	0.09	0.06	0.05
Soil reaction (pH)			
A1 (0–7 cm)	5.1	6.1	6.5
Bt (55–65 cm)	5.2	5.0	5.2
C (115–130 cm)	6.2	6.0	5.2
Base saturation (%)			
A1 (0–7 cm)	90	68	80
E (20–30 cm)	68	80	42
Bt (55–65 cm)	78	87	78
Depth (cm) of leaching of carbonates	90	–	115
Silt-size opal phytoliths (mT/ha) in solum (largely in A)	20	10	5

Source: Based on representative data from Daniels et al. 1962; Jones and Beavers 1964; Riecken 1965; White and Riecken 1955; and Andrews and Dideriksen 1981.

considered in terms of (1) plant communities, (2) individual trees, and (3) particular kinds of organic matter concentrations. In studies of soil genesis we deal with a complex mosaic of overlapping biological communities that reflects not only present, but also past, distribution patterns of biota.

Sequences of soils along transects from grasslands to forests have long interested ecologists (Hole and Nielsen 1970). Transitions from grass-vegetated soils with their thick A horizons to forest-vegetated soils with thinner A horizons underlain by light-colored E horizons (Mollisols and Alfisols, respectively) have been carefully studied in Iowa and Illinois. Notable contrasts were found in ratios of clay content ratios from B to A horizons, higher N content to greater depths, and more opal phytoliths in grass-vegetated soil (Table 3.7). Most, but not all, grasslands developed in or near ustic or xeric soil moisture regimes. The usual assumption is that at the onset of dryness fires, set by aborigines and/or lightning routinely maintained the prairie by eliminating invading trees and fertilizing the prairie grassland ecosystem with plant available inorganic forms of essential elements in the ash. As humans cultivated large areas of grass-vegetated land, fires were largely eliminated; trees invaded uncultivated area that had been vegetated by grass causing marked changes in soil characteristics (Dormaar and Lutwick 1966). The demise of grassland vegetation causes decomposers (invertebrates and microorganisms), which depended on fibrous root debris from the grasses, to focus on stored humus (Jenny 1980). As a result, a bleached horizon (E) appears under the advancing forest in the formerly entirely black A horizons formed under grass vegetation. Soil solution studies in Wisconsin showed that forest growth enhanced podzolization processes in sandy soils formed under grassed prairie vegetation and cultivated for some years (Quideau and Bockheim 1996).

In the Jemez Mountains of north-central New Mexico, Allen (1984) observed mollic epipedons 30- to 80-cm thick under mountain grassland on south-facing slopes, even under recent invasions of pine and spruce (*Picea engelmanni*) and aspen (*Populus tremuloides*), and with thin, about 10-cm thick A horizons under old-growth forest on north-facing slopes.

Bockheim (1972) reported a biosequence from a subalpine Krummholz forest to meadow, in which corresponding soils showed albic and umbric horizons, respectively (Ugolini and Edmonds 1983). Ugolini et al. (1981) described a white spruce (*Picea glauca*) forest and adjacent alpine tundra (lichen-mixed heath) in the central Brooks Range, Alaska, in which a single kind of soil lay under both ecosystems. No explanation is forthcoming for this anomaly.

A unique biosequence of soils exists at the San Dimas Experimental Forest in southern California. In 1946, monocultures of several chaparral shrub species and Coulter pine (*Pinus coulteri*) were planted on large (5.3 × 5.3 m horizontally and 2.1 m deep) lysimeters that had been previously filled with a homogenized and uniform sandy loam parent material. Thus, at their inception all soil-forming factors except vegetation were constant for these soils. By 1959 different microclimates had developed as a function of plant species, with cooler soil temperatures under the dense canopy of scrub oak than under the more open canopy of pine (Qashu and Zinke 1964). By 1987 obvious soil morphological differences could be observed. Earthworms that thrived under the scrub oak served to mix organic matter into the mineral soil, create water-stable aggregates, thicken the A horizon, keep the litter layer relatively thin, and inhibit evidence of eluviation. In contrast, earthworms and other mixing fauna were absent under pine, so a thick litter layer accumulated, the A horizon was thin, and a weak argillic horizon formed (Graham and Wood 1991; Graham et al. 1995). Differences were also found in exchangeable cations (Ulery et al. 1995; Quideau et al. 1999), organic carbon dynamics and sequestration (Quideau et al. 1998, 2000, 2001; Feng et al. 1999), and weathering rates (Quideau et al. 1996).

Effects of Single Trees on Soils. There are notable instances of spatial soil property relationships to single trees. Bodies of Spodosols are present under kauri pine trees and in New Zealand are called "basket Podzols" because they are that shape in plain view of the tree canopy. They are thickest next to the tree where acid bark accumulates <2-m thick, and they thin outward to the drip line (Swindale 1955). However, no

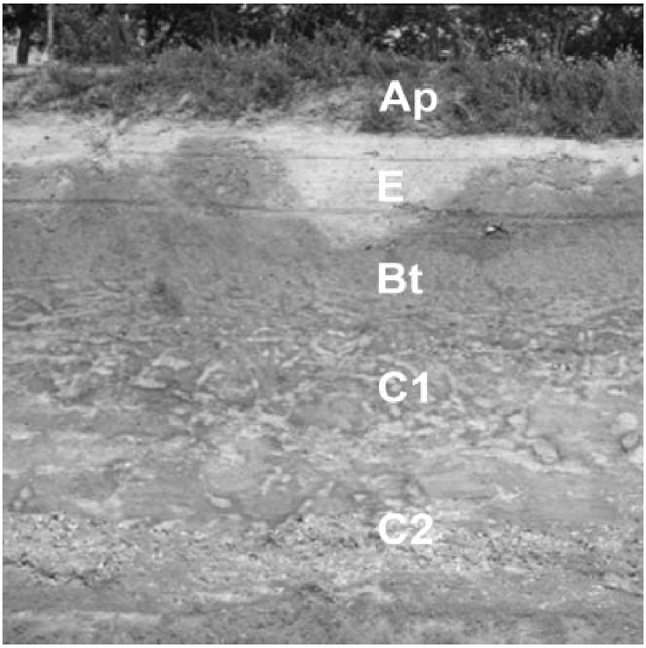

Figure 3.14. Road bank exposure of profile variations caused by uprooting of a tree. Site in Eastern North Carolina where an uprooted tree moved kandic horizon material, right to left with deepened E horizon at former tree site. All wood has decomposed. For color detail, please see color plate section.

alteration was noted under single digger pines *(Pinus sabiniana)* scattered in grassland soils (Xeralfs and Xerochrepts) in the Sierran foothills of California (Jenny 1980).

During less than 3,000 years of growth, a single specimen of sequoia *(Sequoiadendron giganteum)* in the Sierra Nevada Mountains of California has heaved and pushed the original soil horizons out of all recognition as the trunk (8 m in diameter) formed (Jenny 1980). But 3 millennia is ample time for new horizons to form that are characteristic of the Dystrudepts that exist today (Zinke and Crocker 1962; Hole 1968). Depth and radial (lateral from a tree trunk) functions of the soil show some similarity with respect to thickness of O horizons, bulk density, and pH.

Under a single specimen of American beech *(Fagus grandifolia)*, an annular soil pattern develops, with a ring of Inceptisols adjacent to the bole of the tree, where stem flow quintuples the normal precipitation of the region, and an outer ring of Alfisols on which precipitation is more normal (Gersper and Holowaychuk 1970a, 1970b).

Many individual trees are uprooted in windstorms, tearing up the root plate with attached soil. As the plant roots decompose, soil drops to form a mound as much as a half-meter high, adjacent to a pit from which the soil and root mass were extracted. (See Figure 3.14.)

In northern Wisconsin, 350 depressions, called "cradle knolls," caused by uprooted trees have been counted in a single hectare. Rate of production of these features has been about 1% of the land surface per 500 to 1,000 years. Gaikawad and Hole (1961) have described the distortion of soil horizons observed in tree-tip mounds. Schaetzl and colleagues have studied this phenomenon extensively (Schaetzl 1986; Schaetzl et al. 1989).

Animals and Soils. This section is concerned with the effects of animals on soil, with emphasis on those animals that live in the soil. The diversity of soil animal species, even within a small area, typically equals that of a coral reef (Wallwork 1970). All major animal phyla, except Coelenterata and Echinodermata, are represented in soils of the world. Soil animals are not organized into well-defined regional communities, "perhaps because bulk samples include a variety of microsites, and because microsite conditions change rapidly" (Anderson 1977). Soil fauna include ancient forms of terrestrial life (Southwood 1978) that evolved before plate tectonic processes had dismembered the primary land mass. A useful display of 61 sketches of representative endopedonic and exopedonic animals was presented by Hole (1981). Three of these are included in Figure 3.13.

The biomass of soil animals in the upper 30 cm of soil may be about 1 percent on a dry-weight basis but may account for 25 percent of the consumption of oxygen in a forest floor (Barley 1961). Kurcheva (1960) reported that over a period of 140 days, soil animals and microflora reduced *Quercus* leaf litter 5 times faster than microflora alone. The biomass of soil fauna is a nutrient reserve, but not a nutrient source, in itself. Animals determine certain patterns of distribution of materials in soils and have a multiplier effect by concentrating certain elements in their bodies (Helmke et al. 1979). Extinction of an influential species of animals changes fabric and nutrient cycles of a soil. For example, the passenger pigeon (*Ectopestes migratorius*) (Schorger 1973), before its extinction in 1900, migrated in flocks numbering as many as 2 million birds and nested en masse in oak trees in the Upper Mississippi Valley region of North America. The large-scale consumption of insects and mast, and deposition of excreta and nest debris accomplished by this species has ceased, many of the oak trees having been removed by farmers as they prepared their fields for cultivation.

Invertebrates and Soils. In general, the soil fauna that have the greatest impact on soil structure (Lee and Foster 1991) and a variety of other soil properties are earthworms, ants, and termites. In serpentine barrens in Rhodesia, termites improve the Mg/Ca ratio of soil in mounds, allowing grasses to form a dense sward on the mounds (Wild 1975). Nutrient cycling is not just a vertical function, but it is also a radial function to and from the nests, in landscapes in which 1,000–10,000 termites may be present per square meter (Lee and Butler 1977). Termites create cements with which they bind sand grains into the walls of the termitaria. A flagellate protozoan in the gut of a termite decomposes ingested cellulose. Nye (1954) proposed CrT (Cr: creep, T: termite) soil horizons, which are fine textured and about 45-cm thick, composed of particles of clay translocated upward bit by bit by the insects. The whole layer gradually creeps (Cr) downslope under the pull of gravity.

Ants "are the chief agents for introducing organic matter into the soil in tropical rain forests" (Weber 1966) and do their work in close coordination with termites. The western mound-building ant (*Formica cinerea*) of the grasslands of central North America makes mounds 40-cm wide at the base and 30-cm high in well-drained soils (Baxter and Hole 1967), but 150-cm across and 40-cm high in poorly drained, clayey soils (Denning et al. 1973). In the latter, macropores occupy 16 percent of the upper portions of the mounds, accounting for a saturated hydraulic conductivity of 10,000–44,000 cm day^{-1}, which is 1–4,000 times the maximum precipitation in a 24-hr period in the (Wisconsin) region. Lyford's study (1963) of ants that make small (1-cm high by 7-cm wide) surficial deposits of soil showed that the upper 35 cm of soils in the Harvard Forest have been homogenized by these insects. Wang et al. (1995) estimated that mixing by the ant *Lasius neoniger* resulted in a turnover of the upper 30 cm of soil several times during 1,000 to 2,800 years in Wisconsin.

Earthworms occur on every continent except Antarctica and comprise over 3,500 species. They can have major effects on soil microbial, chemical, and physical properties. The nature and location of these effects in the soil is closely linked to three basic ecological strategies. Epigeic species live in O and upper A horizons where they burrow laterally and, through feeding and casting, mix litter and mineral soil material, promote microbial colonization, and speed organic matter decomposition. Anecic species have relatively deep vertical burrows into which they pull surface litter, thereby mixing organic material into the subsoil. They cast on the surface, mixing mineral material into the litter. Endogeic species burrow within the mineral soil, feeding on soil organic matter and microbes associated with the rhizosphere (Hendrix 2000).

Earthworms have been found to degrade the structure in some soils (Thorp 1949; Nooren et al. 1995) and greatly improve it in others (Van Rhee 1977; Blanchart 1992). As soil passes through the digestive tract of an earthworm, interparticle bonds are disrupted. Thus, fresh earthworm casts are highly dispersible and may contribute to soil erosion. Drying and aging increases the aggregate stability of the casts (Shipitalo and Protz 1988). By consuming leafy forest floor material, *Lumbricus terrestris* exposes soil to impact of raindrops and canopy drip (Nielsen and Hole 1964).

Work of Wiecek and Messenger (1972) indicates that if one dropped acid at random on a deciduous forest floor in northern Illinois, small explosive events of effervescence would take place because of the presence of numerous biogenic calcareous nodules, about 1 mm in diameter, that are released by earthworms, particularly through death and decomposition. About 1 kg of these carbonate particles are contributed annually by *Lumbricus* to each hectare in such a hardwood forest. Substantially larger pores, in the form of earthworm channels, are present in epipedons under no-till management as compared with conventional tillage (Drees et al. 1994). Nye (1954) described a *CrW* (creep worm) soil horizon, about 2.5-cm thick that consists of granular peds formed by worms. This horizon slowly moves (creeps: *Cr*) downhill under the influence of gravity. In Nigeria, some earthworm casts described by the same observer stand 6-cm tall and are 1.5 cm in diameter. Darwin early (1896) observed the role of earthworm casting in burying the land surface (including Roman roads).

Cicada (*Magicicada* spp.) nymphs tunnel through soil and backfill the voids, creating thumb-size peds (Hugie and Passey 1963). These backfilled burrows are preserved in paleosols of the Palouse loess in eastern Washington and can be used as indicators of paleoenvironmental conditions (O'Geen and Busacca 2001). Trap-door spiders (Araneida), wasps (*Bembix* spp.), Coleoptera beetles, grasshoppers (*Schistocera* spp.), mole crickets (Gryllotalpidae), and bees (*Megabombus* spp.) are among the many insects that burrow in soil. Dung beetles, including the giant *Heliocopris dilloni* that disposes of elephant dung in Kenya, bury droppings of many kinds of animals (Breymeyer 1974). Crayfish (*Cambarus* spp.) build cylindrical chimneys on the surface of soils that have fluctuating water tables (Thorp 1949). Their backfilled burrows (krotovinas) are major morphologic features and impact hydrologic behavior in some such soils (Vepraskas and Wilding 1983).

Vertebrates and Soils. Moles (Talpa), marmots (Marmota), badgers (Taxidae), prairie dogs (Cynomys), pocket gophers (Thomomys), and arctic ground squirrels (Citellus undulatus) are among the many vertebrate burrowers in soil. Pocket gophers are particularly effective at mixing the upper several decimeters of soil as they search for food. This can have the effect of thickening epipedons (Laurent et al. 1994). Besides moving tons of soil each year (Thorp 1949), burrowing mammals move air underground. Vogel and Bretz (1972) documented this effect of prairie dog burrow geometry, reporting that differential wind velocities of 20 cm sec^{-1} and 10 cm sec^{-1} at upper and lower entrances, respectively, would replace air in the burrow every 10 minutes. A filling, usually of dark surface soil, of an abandoned animal burrow is called a krotovina (a Russian term meaning mole hole).

Biogenic mounds are not to be confused with eolian, periglacial, gilgai, and tree-tip mounds (Slusher 1967; Thornbury 1965; Oakes and Thorp 1950; Troedsson and Lyford 1973) nor with Indian mounds and raised vegetable beds in prehistoric gardens.

Some birds, notably petrels (*Pachyptela* spp.) and shearwaters (*Puffinus* spp.), burrow and nest in soil. The unusual thermometer birds (*Leipoa ocellata*) maintain optimum temperatures in heaps of fermenting leaf litter by manipulating them for proper incubation of the eggs laid in them (von Frisch 1974). Ornithogenic soils in Antarctica are those occupied by Adelie-penguin (*Pygoscelis adeliae*) colonies that contribute guano, eggshells, feathers, and bone fragments to the substrate (Syroechkovsky 1959).

Humans and Soils. The work of humankind (Bidwell and Hole 1965; Yaalon and Yaron 1966) in cultivating fields, removing hills, filling in low places, irrigating dry soils, draining wet soils, reducing (or increasing) fertility of soil, and accelerating or reducing erosion and deposition through surface water manipulation is familiar to us.

People have converted "wild soils" (Hole 1974; Bouma and Hole 1971) to "tame" or "domesticated soils" more capable of serving their needs for food and fiber production.

Soil Taxonomy attempts to classify both "wild" soil and the "domesticated" soil by the same name when alteration does not exceed expected cultural practices. However, where humans have made extensive soil modifications, characterized by the presence of fragments of diagnostic soil horizons that were broken up by deep plowing, leveling, and construction activity, separate categories have been established. Hundreds to thousands of years of paddy (flooded) rice cultivation in China has created specific surface conditions that has prompted Chinese soil scientists to define specific epipedons and near surface features in their system of soil classification (CRG-CST 2001). Briefly these are Siltigic epipedon (accumulation of silt from application of silt-rich irrigation water), Fimic epipedon ('mellowing' from long-term additions of manure, organic refuse, etc.), Anthrostagnic epipedon (results from submerged cultivation with formation of a plow pan), and the Hydragric horizon (accumulation of iron and manganese oxides under anthrostagnic epipedons). Specific chemical and physical parameters and depths are specified in their definitions.

Although localized areas of extensive excavations and/or deposition of soil material by humans are readily seen, it is their cultivation of approximately 11 percent of the earth's land area for the growing of human food that is most extensive. Perhaps the most extensive alteration of soils by humans is their harvest and transport of life-essential elements in the food they consume and their practices of recharging those elements by applying chemical and organic fertilizers to areas of continued food crop production.

In Figure 3.15 the flow of elements is schematically traced through mineral, organic, and available pools in the soil and then combined with C, H, and O from air and water in human food plants. As parts of the food plants are harvested, some plant parts are deposited to be incorporated into the organic pool at the site of crop growth, but the most essential element-rich parts (usually the seed) are transported for food at sites of human habitation titled 'city' in Figure 3.15 but considered as small as an individual farmer's house. Whether humans consume these elements as vegetable or animal products, those elements contained are seldom totally returned to the site where the food was grown. Among the essential elements in human food plants, only C, H, and O are directly obtained from air and water. The other essential elements enter plants as inorganic ions in the available pool. Small amounts of nitrogen enter the soil in precipitation, but most is naturally obtained from the air via nitrogen-fixing microbes in the soil and becomes available to plants as inorganic NO_3^- and NH_4^+ ions only after the organic compounds are decomposed. All other essential elements originate in soil minerals and enter the available pool as inorganic ions as minerals decompose or are recycled via decomposition of organic material. There is only a finite quantity of minerals bearing these elements in a given soil. As human extraction of food depletes these fixed quantities, the growth rate of food plants declines. Over generations humans have learned to fertilize areas and thereby enhance and sustain food crop production, first with manures and organic residues added to the organic

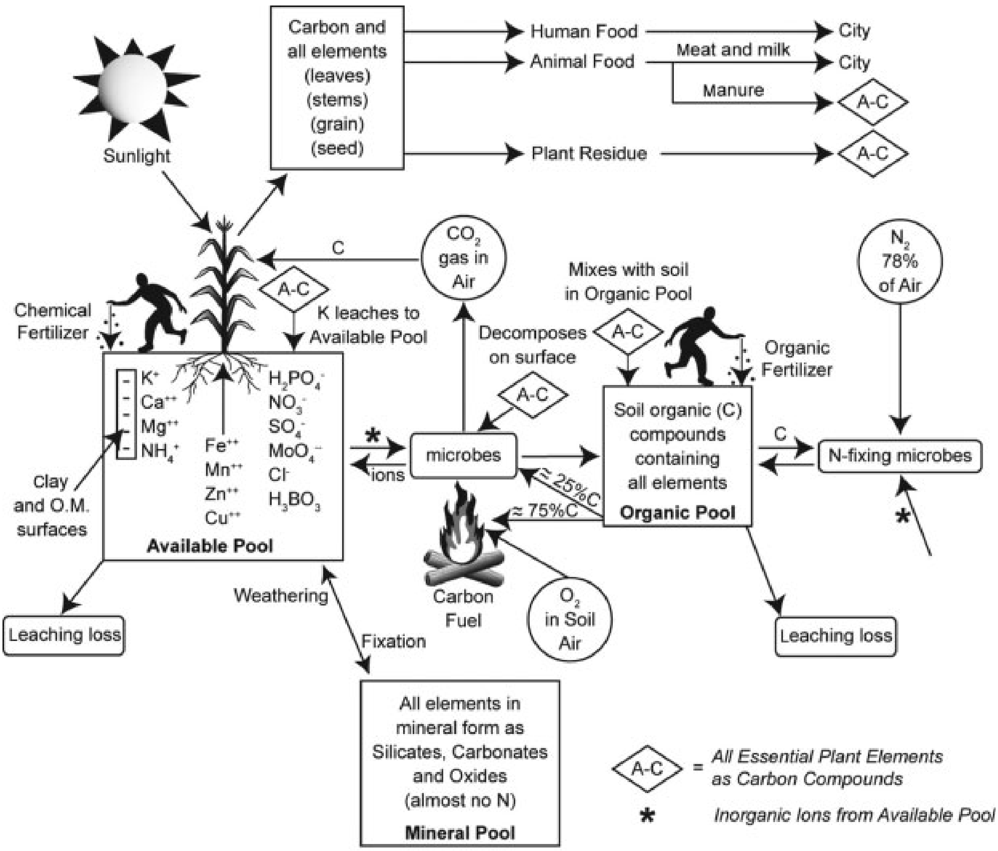

Figure 3.15. Elemental flow in human food production (Source: Buol 2008).

pool and more recently in human history with concentrated chemical fertilizers added to the available pool in the soil.

Human food crops are prodigious consumers of essential elements from the soil. Compare the amounts of N, P, K, and Ca in average yields of corn, rice, and soybeans farmers in the United States attain with the amounts contained in the total aboveground biomass of 22 and 60 year old plantation grown Loblolly pine (*Pinus Teada*) (Table 3.8). More important than comparisons of amounts is the reality that the food crops require those amounts, plus amounts needed for roots in a growing period of less than about 100 days compared to the years required for the pine growth. Although the large biomass of the pine trees may look impressive to the human eye, in reality the amount of essential elements and more importantly the rate at which they are taken up by trees is only a fraction of human food crop requirements. These comparisons illustrate why slash and burn farmers quickly reduce available nutrient supplies with only a few crops, abandon their fields as yields decline where upon trees reestablish, grow for several years on a low rate of available elements and after

Table 3.8. Approximate elemental content of some human food crops and pine trees

	Yield	N	P	K
Crop	kg ha^{-1}			
Corn (grain)	9,072	151	26	37
Corn (stover)	10,080	112	18	134
Sorghum (grain)	3,629	56	12	15
Sorghum (stover)	6,720	73	10	88
Rice (grain)	4,032	56	10	9
Rice (straw)	5,600	34	6	65
Soybean (grain)	2,688	168	18	52
Soybean (straw)	3,360	20	2	22
Pine (20 years)	84,672	180	19	90
Pine (60 years)	224,000	256	24	194

Source: Anonymous 1972; Tew et al. 1986.

sufficient growth can again be cut and burned to release essential elements in available inorganic form to provide for one or two food crops.

Where external sources of fertilizer are applied to human food crops, the chemical composition of the soil is greatly enhanced, and production of food crops per unit of land is greatly increased. Where external enrichment is not available, the global condition for much of human history that continues today, in many parts of the world vigorous food crop growth rapidly declines. With decreased vegetative cover, soils are more exposed to erosion. As stated in the *Soil Survey Manual* (Soil Survey Staff 1951, p. 261), "In fact, if continuing cultivation is assumed, generally, although not always, low fertility can be regarded as a main cause of erosion." Thus, as with other organisms, the influence of humans on soil formation and properties is hard to quantify but clearly the association of erosion with early human efforts to grow food crops without access to external sources of essential elements stems from nutrient depletion.

Time as a Factor in Soil Formation

During soil formation additions and losses of material alter the vertical location of the soil. In standard dictionaries, space-time is defined as a continuum in which all things exist. Soil occupies that space below the land surface that is capable of supporting the root systems of plants. Soil is observed at a given time, and its vertical profile is described and sampled with reference to the surface boundary. Most studies tend to accept that the soil surface is a fixed point, but in reality over time, the vertical location and thus space observed as soil below moves. Some of the vertical losses and gains are relatively rare episodic events of landslide, major erosion, and flooding that completely remove or bury existing vegetation and soil. Following such events, plants again populate the new land surface, and a new volume of material becomes the space we define and study as soil related to the vertical datum point of the soil surface. Where depositional events bury a preexisting soil that we expose in examining the present soil, we identify buried soil horizon(s). (See Chapter 2.)

Most erosion and depositional events take place without significantly affecting existing vegetation. Other processes, like dissolution of soil minerals by percolating water, creep on hillsides, and depositions from aerosol dusts are so slow they easily escape human observation and measurement.

The degree to which materials within soil are altered depends upon the amount of time each material experiences a specific process that causes its alteration and the resistance that material has to alteration. The rate of change that will occur can be expressed by the equation $d\mathrm{S} = \mathrm{T} \times (\mathrm{I} - \mathrm{C})$ where dS = the rate a soil property changes; T = time; I = the intensity exerted by the process to change that property; and C = the capacity of the material in the soil to resist change. Where intensity of a process to affect change (I) is much greater than the capacity of the material to resist change (C), the rate of change (dS) is rapid. Where the capacity of the soil material to resist change (C) is nearly equal to the intensity of the physical and biogeochemical processes to cause change (I), the rate of change is slow. Where (C) and (I) are equal, or where (C) exceeds (I), the soil property does not change with time.

Concept of Time Zero in Soil Formation. Time zero in soil formation is a point in time when a catastrophic event is completed, and sufficient new material is exposed at the land surface. Such an event may be significant deposits of sediment, loess, volcanic ash, colluvium, lava flows, or land forming by humans that completely destroys preexisting soil and vegetation. Less abrupt changes in topography or water-table depth caused by geologic uplift or the rapid retreat of a hillside due to geologic erosion (Ruhe 1960; Thwaites 1956) may also be considered to mark time zero. Events that clearly define a time zero for soil formation are relatively rare. Most soils we observe have undergone a series of less drastic changes each of which may have been significant in altering its composition. Such changes include burning existing vegetation followed by a natural succession of vegetative communities or the introduction of agricultural crops, with attendant cultivation practices and later abandonment. Unknown historical changes in vegetative- and/or human-induced cultivation practices may create persistent soil properties although their presence is obscure to observations of current environmental conditions.

Spatial Changes of Soil Over Time. Evaluation of time as a factor in soil formation often assumes that the surface of the soil remains fixed and thus the volume (depth) examined has remained the same over the time span under study. In reality several reactions that either lower or raise the soil surface take place during time frames in which soil-forming processes are evaluated. Perhaps the most easily understood process of spatial change is the lowering of the soil surface via erosion. Of equal, and

Table 3.9. Dissolved and solid removal rates within major watersheds in the United States

Drainage Region	Dissolved ($T\ mi^{-2}\ yr^{-1}$)	Solid ($T\ mi^{-2}\ yr^{-1}$)	Total ($T\ mi^{-2}\ yr^{-1}$)	Total Rate of Removal[a] (in. 1000 yr^{-1})	Total Rate of Removal[b] ($yr\ cm^{-1}$)
Colorado River	65	1190	1255	6.5	61
Pacific Slopes (Calif.)	103	597	700	3.6	109
W. Gulf (Texas)	118	288	406	2.1	187
Mississippi	110	268	378	2.0	197
S. Atlantic and E. Gulf	175	139	314	1.6	246
N. Atlantic	163	198	361	1.9	207
Columbia River	165	125	288	1.5	262
Total U.S.A.	121	340	461	2.4	164

Source: Judson and Ritter 1964.
[a] Calculated as uniform removal from the surface of all the land in the river watershed.
[b] Transformed for compatibility with text.

in some places, greater magnitude is the lowering of the land surface by dissolution of minerals in the soils and underlying parent material.

Judging by material in solution in rivers, Clarke (1924) concluded that the land surface of the earth is being lowered 30 cm every 30,000 years (1,000 yr cm^{-1}). Judson and Ritter (1964) calculated dissolution and erosion removal rates within major watershed areas in the United States from records of dissolved and sediment loads in rivers (Table 3.9). Assuming uniform removal from the total land area within each watershed, they calculated the rate at which land surfaces would be lowered by erosion of solid soil particles and leaching of dissolved elements. Summarized for the entire United States, it was calculated that it would take 164 years to lower the land surface 1 cm, 631 yr cm^{-1} by dissolution and 222 yr cm^{-1} by erosion of solid particles, a somewhat faster rate than calculated by Clark.

Major watersheds differ in both total rate of loss and proportion attributed to mode of removal. Solid removal accounts for nearly 95 percent of the loss in the mountainous, mostly arid and sparsely vegetated Colorado River basin and only about 44 percent of the loss in the more humid Columbia River watershed and South Atlantic and Eastern Gulf areas of the United States. Solid removal loss rate is calculated as 559 years cm^{-1} in the southeastern United States while only 64 years cm^{-1} in the Colorado River watershed. Dissolution loss rate in the humid southeastern part of the United States is calculated as 447 years cm^{-1} while in the Colorado River watershed, the dissolution loss is 1,220 years cm^{-1} of soil. Much of the dissolution losses are likely to take place as weatherable primary minerals in consolidated parent rock dissolve to form saprolite and ultimately soil material. Dissolution of rock to form saprolite is known to be a reduction in bulk density, as shown in Table 3.2, and may not result in actual lowering of the land surface. Much of the solid removal is via

Table 3.10. Erosion rates of solid particles within individual river basins

River Basin	Location Measured	Erosion Rate (in. 1000 yr^{-1})	Erosion Rate (yr cm^{-1})[a]
Delaware	Trenton, NJ	0.8	492
Rappahannock	Remington, VA	0.7	562
Tombigbee	Jackson, AL	0.5	787
Alabama	Claiborne, AL	0.5	787
Rio Grande	San Acacia, NM	1.8	219
Pecos	Puerto de Luna, NM	3.6	109
Mississippi	Baton Rouge, LA	1.3	303
Colorado	Grand Canyon, AZ	5.6	70
Sacramento	Sacramento, CA	0.5	787
San Joaquin	Vernalis, CA	0.1	3937
Eel	Scotia, CA	30.4	13
Mad	Arcata, CA	19.3	20
Snake	Central Ferry, WA	0.7	562
Columbia	Pasco, WA	0.5	787

Source: Judson and Ritter 1964.
[a] Transformed for compatibility with text.

erosion from the soil surface with some portion coming from river channel deepening and bank erosion. It can be reasoned that in areas where dissolution loss rates exceed solid removal rates, soil profiles will be thick and where the removal of solids is more rapid than dissolution loss, soil profiles will be less thick.

A more detailed analysis of solid particle erosion rates within the watersheds of individual rivers reveals considerable variation among local watersheds within each region (Table 3.10). It is clear that the greatest erosion rates are in more arid and mountainous regions where only a small proportion of the watersheds are cultivated. The high rates of erosion within the primarily forested watersheds of the Eel and Mad rivers in Northern California result from tectonic instability and the rapid erosion that follows. Intensely cultivated level land, for example the San Joaquin River basin, has a very low erosion rate.

Although the averaging of soil loss over entire watersheds in Tables 3.9 and 3.10 provides a datum for considering the space-time relationship to soil formation, extreme unevenness exist spatially within watersheds. Trimble (1999) reported erosion losses from upland areas and depositional gains on floodplains and losses to the Mississippi River in the 88,920-acre (360 km^2) Coon Creek basin in west-central Wisconsin. Coon Creek watershed is a rather hilly area with rather broad floodplains adjacent to the main channel and major tributaries. Farming began about 1850, and by 1930 severe erosion on farmland prompted federal- and state-sponsored soil scientists to initiate intensive and long-term studies to determine the effectiveness of soil conservation practices on rates of erosion. As farmers implemented soil conservation practices and land use changed, impressive reductions in erosion and depositional

Table 3.11. Average annual movement of soil material in Coon Creek watershed in three eras

	Erosion Sites			Deposition Sites		
Time Period	Upland Sheet Erosion	Upland Gullies	Tributaries	Tributary Floodplains	Main Stream Floodplain	Mississippi River
1853–1938	359*	80	46	138	309	42
1938–1975	126	71	68	42	183	40
1975–1993	84	21	23	28	61	41

* All Units are 10,000 tons.
Source: Trimble 1999.

rates were observed within the watershed (Table 3.11). Discharge of sediments into the Mississippi River remained nearly constant despite the impressive reduction of erosion on upland sites. Source of the sediment loss to the Mississippi River was attributed to stream bank erosion.

Localized erosion and depositional patterns in the Coon Creek watershed can be explained by an examination of the erosion process(s) involved. Most erosion takes place during short periods of intense rainfall. On sloping sites soil particles are detached from the soil surfaces by the energy of raindrop impact and suspended in water rapidly flowing downslope. During and immediately after the rain event, sediment-bearing water overflows the stream channels and covers the adjacent floodplains where the velocity of flow diminishes and sediment is deposited. In most cases existing vegetation is not completely buried by these depositions and soon recovers to grow anew on the now higher surface of the soil. Likewise on the eroded sites vegetation either remains in place or quickly reseeds and regenerates on a lower soil surface.

No two watersheds have the same configuration, and the Coon Creek observations are certainly not applicable to steeper land wherein the floodplains are narrow or almost nonexistent and streams flow with greater velocity. In such watersheds the proportion of eroded material reaching major rivers or lakes would be much greater than measured in Coon Creek. However, this study clearly shows that within a limited spatial location, the soil surface we currently observe has moved either up or down during a timeframe in which soil-forming processes have actively created specific soil features.

A Conceptual Continuum of Space and Time. With recognition that the datum point for examining soil, that is, the soil surface, moves over time, we must consider how this movement interacts with the other soil-forming processes in the formation of the soil features we observe. Figure 3.16 is a conceptual representation of the space-time continuum in which soils at a fixed geographical location may change over time. In Figure 3.16, undefined segments of time (t) are represented on the horizontal axis, and segments of vertical space (e) are plotted on the vertical axis. Within that vertical space below the soil surface are two types of geologic materials,

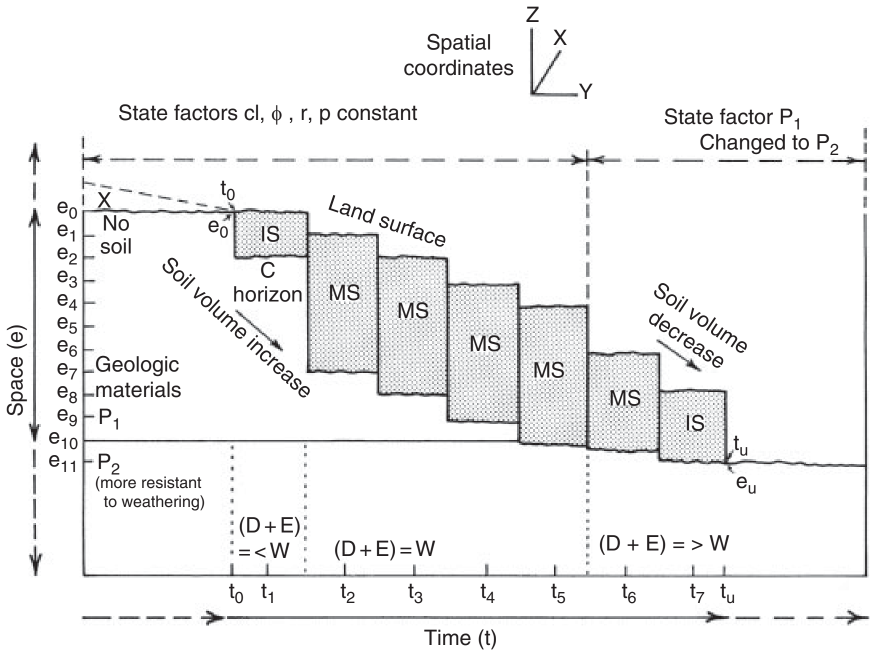

Figure 3.16. Diagram of the space-time continuum (*e-t*), in which any given soil body exists from its time$_{zero}$ (t_0) and space$_{zero}$ (e_0) to its time$_{terminus}$ (t_u) and space$_{terminus}$ (e_u). The *shaded area* represents a spacetime sequence (chorochronosequence) of selected soil profiles (in outline) to illustrate spatial and temporal changes in the course of formation and extinction of a soil body. IS = immature soil; MS = mature soil; *D* = rate of dissolution processes; *E* = rate of erosion processes; *W* = rate of soil formation processes. Wedge X (*upper left*) represents a cover of glacial ice that prevented soil formation before $t_0 - e_0$.

P_1 is an easily weathered material and P_2 is an inert and nearly impermeable material that resists weathering and soil formation. A series of soil bodies exist between units in vertical space, e_0 to e_u and units of time, t_0 to t_u. Prior to t_0, no soil was present. Although it is not critical to understanding the concepts presented, we consider t_0 to mark a catastrophic even such as a landslide or the retreat of glacial ice that exposed soil material capable of supporting plant growth at a land surface and thus soil formation. Also, the site selected is on a gentle slope subject to both erosion and leaching.

From time t_0 through t_1, a thin soil with an A-C horizon sequence (considered immature [IS] because it lacks a Bt horizon) is formed as plants strike their roots into the soil and add organic carbon to form an A horizon as they die. Weathering and processes of soil formation (W) alter the minerals present at a more rapid rate than erosion (E), and dissolution (D) altered the vertical position of the soil surface, that is, (D + E) <W. During time between t_1 and t_2, the soil thickens and develops an A-B_t-C horizon sequence characteristic of the area and considered a mature soil (MS). Some vertical movement of the soil has taken place as erosion (E), and dissolution (D) lowers the soil surface in space (e).

After attaining the morphology and composition of a mature soil, the soil is seen to maintain the same characteristics between times t_2 and t_5 as erosion (E) and dissolution (D) lower the position of the soil at the same rate as weathering and soil formation processes (W) altered underlying material, that is, (D + E) = W. This conceptual equation is an expression of a steady-state reaction in which parent material is altered by physical and biogeochemical processes at the same rate as soil material is removed from the site. Soil characteristics remain nearly unchanged during this time interval, and the soils in this sequence are considered mature (MS) but moving downward in space into relatively easily weathered parent material (P_1). At some time after t_5, the lower extent of weathering and soil formation encounters the more resistant parent material (P_2). At that time weathering and soil formation processes are not able to vertically progress at the same rate as erosion and dissolution lower the soil surface. During some time near t_6, the soil may have an A-Bt-C horizon sequence and be considered mature, but the A and B horizons have formed from material that differs from P_2 and the horizon sequence is more correctly designated A-Bt-2C. (See Chapter 2.)

From time t_6 to t_7, soil thickness decreases to such an extent from erosion and dissolution losses, that is, (D + E) >W, that A horizon features envelope the former Bt horizon and the soil, now with only an A-C horizon sequence is considered immature (IS). At some time after time t_7, if erosion and dissolution rates continue to exceed weathering and soil formation no longer penetrates the resistant parent material (P_2), soil may no longer exist, as represented by a terminus of soil in both time (t_u) and space (e_u) and a non-soil outcrop of P_2 is exposed at the site.

Space-time diagrams like Figure 3.16 are conceptual models that consider landform changes in the same timeframe as the formation of pedogenic features. Space-time models can be constructed to better understand spatial distribution of soils on any landscape. For landscapes that receive material via deposition, or surface accumulations of organic material as in the case of organic soil formation (see Chapter 13, Histosols) the erosion term (E) is considered negative, and if it exceeds (D) the (E + D) value becomes additive with (W) indicating that the space occupied by soil enlarges upward in space (Buol 1992).

In some cases changing soil characteristics related to soil-forming processes have altered the course of soil formation. Some soils in southern Illinois began as moderately well-drained soils undergoing progressive leaching but arrived at a condition of sodium accumulation because of a decline in permeability in the underlying glacial till (Wilding et al. 1963).

Concepts of "Mature" Soil and the Time Factor. Although seldom discussed in recent literature, concepts of soil maturity were prevalent in soil science for many years. Marbut (1928) conceived of a mature soil as one on somewhat sloping land and hence subject to geologic erosion. Others considered level upland as sites of mature soil formation. Nikiforoff (1942, 1949) described the mature soil as "a steady stage of its parent material adjusted to the environment…" "the time factor has no significance after the soil reaches maturity." However, Butler (1958) concluded that changes in soil characteristics continue to occur. He studied soils formed from alluvium of different ages and concluded that soil development continues throughout the history of a soil. He defined the “soil cycle” as the sequence: (1) exposure of a new surface on initial material; (2) soil catenal development; and (3) burial of the catena or its removal by erosion.

The soil that we examine today may meet several possible fates over time:

1. It may continue indefinitely in its current form, with its profile sinking into initial material as fast as erosion removes surface soil and dissolution and leaching remove component minerals (basis for the normal or mature soil concept).
2. It may experience mineralogical changes and become more morphologically differentiated if rates of developmental processes exceed the rate of surface lowering (basis for the “senile” soil concept).
3. It may become the parent material of another soil due to a large and relatively sudden change in climate and/or relief and/or biota.
4. It may be buried by fluvial or volcanic deposits.
5. It may totally disappear as a victim of erosion and its components serve as a parent material for another soil at a new location.

Concepts of age and maturity are no longer directly used as criteria for *Soil Taxonomy*. However, the concept of ‘mature’ soil provides a historical landmark in the evolution of our knowledge of landscapes and soil development that was used as guides for formulation and arrangement of categories in *Soil Taxonomy*.

Absolute Dating of Soil Horizons and Profiles. Radiocarbon (^{14}C) dating methods (Scharpenseel 1971) and fission-track dating (Ward III et al. 1993) are useful to date soils. Several methods of dating soil constituents have recently been reviewed by Cornu et al. (2009).

Most organic carbon additions are in the upper layers of soil. A pattern of increasing organic carbon age with depth is present in soil profiles. In experiments in which ^{14}C labeled organic matter has been added to soils, 80 percent or more of the organic carbon has been found to be lost in a very few years (Jenkinson 1966).

Measurements of the natural radiocarbon at various depths in different soils provide estimates of minimum soil age (Scharpenseel 1972). A significant relationship was observed between organic carbon age and depth in Udolls, Udalfs, and Vertisols (Figure 3.17). In Spodosols, however, because of intense organic percolation and accumulation in spodic horizons, the age-depth relationship was not found to be significant. The Plaggepts (now Plagganthrepts) studied were formed by human additions of organic resides.

Principles of geomorphology and stratigraphy, fortified by radiometric carbon and uranium dates, have been used to determine ages of soils on various geomorphic surfaces (Ruhe 1956a, 1956b, 1969a, 1969b; Ruhe et al. 1967; Parsons et al. 1970;

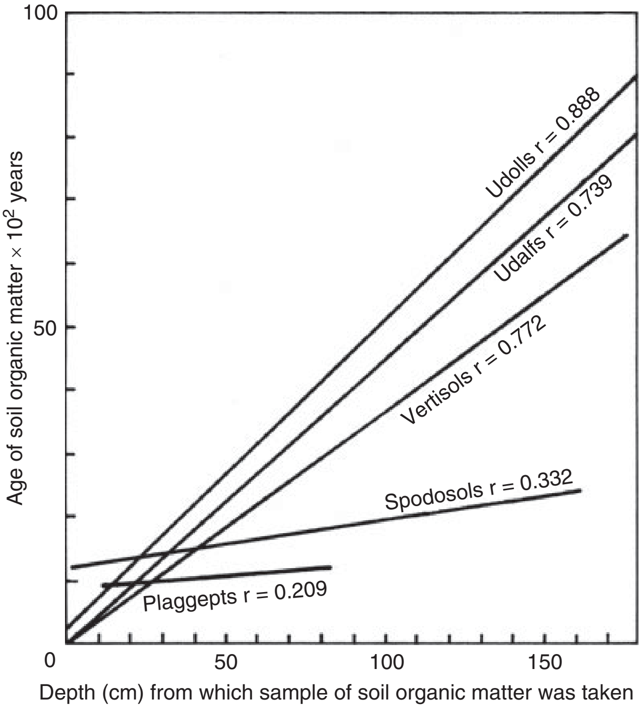

Figure 3.17. Correlation between age and depth of organic matter in some different soils. Source: Scharpenseel 1972.

Parsons and Herriman 1976; Daniels et al. 1971; Jenny 1980; Bockheim, 1990; Daniels and Hammer 1992). A geomorphic surface is a "portion of the landscape specifically defined in space and time" (Ruhe 1969a) or "a part of the land surface with definite geographic boundaries formed by one or more agencies in a given time period" (Daniels et al. 1970). Absolute dating of geomorphic surfaces is not available in many places so dating of them and of the soils formed on them must be relative (Hall 1983) and/or dependent upon evidence provided by stratigraphy and fossil records. For example, in the Rio Grande Valley in southwestern New Mexico, 15 geomorphic surfaces with associated soil properties were identified (Gile et al. 1981). Seven geomorphic surfaces of varying age and soil properties were identified in the Willamette Valley of Oregon (Parsons et al. 1970). These surfaces were Pleistocene to Holocene in age.

In western Iowa soil development and weathering patterns were found to closely correlate with three geomorphic surfaces, varying in age from sixty-eight hundred to several hundred thousand years of age. The oldest soils were at one time buried by loess and now exposed where the loess cover has eroded (Ruhe 1956b, 1969a).

Close relationships were found between soil properties and geomorphic surfaces of varying age in the Middle and Upper Coastal Plain of North Carolina with distinct differences between soils on old stable Pliocene surfaces and Holocene back slope surfaces within a stream valley in the Upper Coastal Plain of North Carolina (Daniels et al. 1970; Gamble et al. 1970a). Soils on the Pliocene geomorphic surfaces have sola more than 2 meters thick, arenic surface horizons and plinthite developed in the lower B horizon. In contrast, soils on the very young Holocene side slopes have sola less than 1 meter thick, lack arenic surface horizons and plinthite, and their clay minerals are not as far advanced in the weathering sequence.

Buol (1965) reported ^{14}C dating of carbonate carbon in carbonate-rich (caliche) layers in Arizona. At a depth of 100 cm, the carbon was 2,300 years old; at 150 cm, 9,800 years old; and below a lithologic discontinuity at 213 cm, 32,000 years old. The conclusion was drawn that the uppermost carbonate-rich layers are related to the current soil and the carbonate-rich layer below the lithologic discontinuity is related to a truncated and subsequently buried soil or carbonate-rich sediment.

Rate of Soil Formation. In discussions of soil losses by erosion, a frequently asked question is, "How long does it take to form an inch of soil?" A popular answer is 100 years per inch of soil (40 yr cm^{-1}). Soil scientists seldom think in terms of inches or centimeters of soil, but rather in terms of horizons, sola, and profiles. Radiocarbon dates of Clarion and Webster soils in glacial drift in central Iowa indicate a rate of A horizon formation of about 8 cm per century (12.5 yr cm^{-1}) (Simonson 1959). Watatuki and Rasyidin (1992) used mass balance calculations of seven elements in parent material, soil, and river waters to calculate that global soil formation rate averaged 178 yr cm^{-1}. They calculated a soil formation rate in specific watersheds of 50 yr cm^{-1} in granite at Hubbard Brook, New Hampshire and 25 yr cm^{-1} in basic pyroclastic parent material in Japan.

Soil formation rate estimates differ because not all soil features form at the same rate, conditions of parent material and energy available for formation differ, and investigators select different soil properties for measurement. Table 3.12 presents some estimates of ages and rates of formation (years per centimeter) of some soil horizons and profiles as examples, but the significance of some estimates is questionable.

Time Factor as Elucidated by Experiments. Laboratory experiments have been useful in elucidating processes of soil formation, such as eluviation, formation of cutans, weathering of primary mica to clay minerals, and transformation of one clay mineral into another. One may conclude that processes observed in the laboratory are relatively rapid and cannot duplicate processes in the soil under natural conditions. Long-term observations of material exposed to soil formation for documented times are rare.

Graham and Wood (1991) examined soil 41 years after placement in lysimeters under pine and oak vegetation in the xeric soil moisture regime of southern California. They found distinct morphologic development related to the interaction of vegetative cover and earthworm activity. Earthworms were active under oak tree vegetation, and

Table 3.12. Some estimates of rate of soil formation

Soil Horizon or Profile	Age upon Completion of Formation (yr)	Depth of Soil (cm)	Rate of Formation (yr cm^{-1})	Literature Citation
Azonal soil (Entisol) on volcanic ash	45	35	1.3	Mohr and van Baren 1954
Hardening of a tropical clay surface soil to literite, following deforestation	35	15	2.3	Aubert and Maignien 1949
A1 horizen of a Vertic Argiaquoll formed in a mudflow near Corvallis, Oregon	133	15	0.1	Forcella 1978
A1 horizons of a Brunizem (Hapludoll) soil formed from weathered loess in Iowa	400	33	12.0	Simonson 1959; Amold and Riecken 1964
2 m thick organic (Histosol) soil in a bog in wisconsin	3000	200	15.0	–
Formation of a Podzol (Spodosol) soil in sandy glacial drift with 10 cm organic and 10 cm E(A2) horizons	1200	57	21.0	Tamm and Östlund 1960
A1 (mull) horizon in a Gray-Brown Podzolic (Hapludalf) soil formed from weathered loess in Wisconsin	265	7	38.0	Nielsen and Hole 1964; Van Rooyen 1973
Solum, including a textural horizon, or a Gray-Brown Podzolic (Hapludalf) soil formed from weathered loess in Iowa	4000	100	40.0	Arnold and Riecken 1964
Decalcified loess in southern Wisconsin	8000	100	80.0	Robinson 1950
A1-E2 horizon sequence in a Gray-Brown Podzolic (Hapludalf) soil formed from weathered loess in Iowa	2500	30	83.0	Parsons et al. 1962
Solum of a Red-Yellow Podzolic (Ultisol) soil in Australia	29,000	300	97.0	Butler 1958
1 m thick solum of a tropical soil (Oxisol in Africa)	75,000	100	750.0	Aubert 1960

their activity enriched surface horizons with clay. Earthworms were absent under pine tree vegetation, A horizons lost clay, and argillic horizons were present. Clearly, the rates at which specific soil properties form are related to the interaction of soil-forming processes that may differ over time in response to succession of vegetative cover.

Chronosequences. A chronosequence is a group of soils for which all soil-forming factors except elapsed time of soil formation are approximately equivalent (Jenny 1941). Study of chronosequences is done by "selecting landscape positions that have comparable state factors, save age," or "splicing together sites that are set apart in space and time" (Jenny 1980). Among the best-documented chronosequences are studies of mudflows in northern California (Dickson and Crocker 1954) and a terrace sequence in Spain (Dorronsoro and Alonso 1994).

It is exceedingly difficult to establish specific ages of soils and ensure comparability of all the other factors of formation. When plants first establish in mineral material they ingest carbon from the air and entrap nitrogen and other essential elements in their tissue. These elements are added to the upper layers of the soil as plant tissue dies. When weatherable minerals are exposed to organic acids, temperature and moisture regimes present near the land surface and not experienced while they were buried deep under the surface alter to secondary clay minerals and clay contents increase. Gains of organic carbon, nitrogen, and clay contents can be measured in initial exposures, but they approach a steady state as microbial activity decomposes the organic residues and returns the carbon to the air, and the supply of weatherable minerals is exhausted. Biotic communities at a given site undergo changes in response to a variety of factors during spans of time that are relatively brief compared to times of soil formation. Chronosequence studies are informative but often site specific.

Perspective

It is from an amalgamation of the soil-forming factors that concepts of spatial patterns of soils are best formed. Experienced soil scientists routinely use their understanding of relationships among the factors of soil formation to predict the kinds of soil they will find on each portion of the landscape they traverse in the process of making a soil map. First, armed with information about climatic conditions and parent material composition in an area, experienced soil scientists will observe the various landscape positions about to be traversed. Where possible, observations are then made of vegetative patterns—both natural patterns in undisturbed areas or in cropped fields. Often these patterns are visible on aerial photographs. With these observations, predictions of what soil conditions to expect are formed. The experienced soil scientist will then traverse the area to verify, via auger or pit examination of soil profiles that predicted soil differences are indeed present. Each observation must be evaluated to determine that it is representative of the area and not a very localized result of disturbance by such things as animal burrows, tree-throw, human disturbance, and so on, in which case additional observations are made.

Consideration of entire soil systems as they spatially exist rather than observations limited to single profiles or pedons provides a better understanding of soils and their role in the total ecology or land use in an area. An understanding of the easily seen current topography and vegetation, an understanding of how these conditions may have changed over time coupled with knowledge of the climatic conditions over the soil and parent material in which a soil has formed, are fundamental to fully understand and appreciate the role of soil in natural and human use ecosystems.

Soil Materials and Weathering

4

Most soils are formed from rock weathered in place or from sediments that ultimately originated from rock. So, the inorganic phase of most soils consists of rock materials plus their weathering products. Since 90% of the earth's crust is composed of silicate minerals, they are the most important components of the soil inorganic phase.

This chapter is devoted to inorganic soil materials. Discussion of soil organic matter processes can be found in Chapter 13, Histosols, and Chapter 15, Mollisols.

Weathering

Weathering is the chemical and physical alteration of rocks and minerals at or near the surface of the earth. The alterations occur because the rocks and minerals are not in equilibrium with the temperature, pressure, and moisture conditions of their current environment (Birkeland 1999). This results in disintegration of the rocks and decomposition and/or modification of both primary and secondary minerals to more stable forms (lower free energy) in their environment. *Primary minerals* are those formed at high temperatures and/or pressures in igneous and metamorphic rocks, whereas *secondary minerals* are those formed at the low temperatures and pressures prevailing at or near the earth's surface in sedimentary rocks and in soils (Jackson 1964a). This definition places an emphasis on the processes and environment of formation, not on composition. Thus, quartz could be primary or secondary, depending on how it forms. Although it is convenient to distinguish between chemical and physical weathering, it is important to keep in mind that most weathering processes in soils are strongly influenced, if not outright controlled, by biological agents (Banfield et al. 1999).

Physical weathering is the disintegration of rocks without any chemical or mineralogical changes but simply by breakup of the rocks into smaller fragments. The main causes of this breakup are stresses within the rocks from day-to-night temperature fluctuations or rapid heating by forest fires, pressure from freezing of water in small voids in the rocks, pressure caused by the growth of salt crystals from saline solutions in cracks, root pressure from plant growth (Birkeland 1999), and release of overburden pressures on rocks by erosion of overlying material. These processes may have a significant direct effect on soil formation by increasing the porosity and surface area of rock bodies and making them more accessible and

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

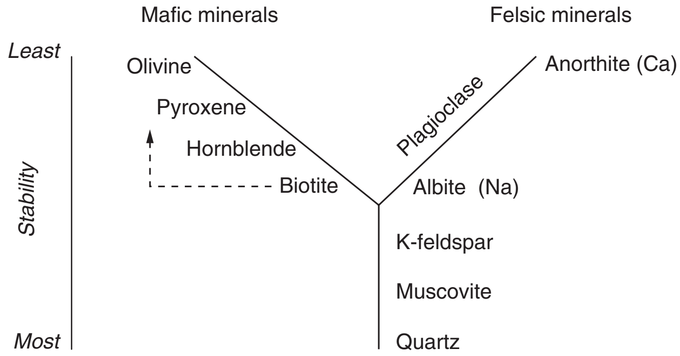

Figure 4.1. Weathering sequence described by Goldich (1938) showing the most easily weathered minerals at the top and the most resistant minerals at the bottom. Goldich's sequence is modified here to show that the simple transformation of biotite to vermiculite occurs at least before hornblende begins to weather (Nettleton et al. 1970), although more extensive alteration occurs later as originally described.

susceptible to chemical weathering (Eswaran and Bin 1978a, 1978b, 1978c; Nettleton et al. 1970; Birkeland 1999). Chemical weathering often weakens rock and mineral structures, allowing them to be more easily disrupted by physical weathering. An example of this is the initial weathering of biotite in granitic rocks. The interlayer K in biotite is relatively easily replaced by hydrated cations, particularly Mg, resulting in a 30–40% expansion of the biotite structure. This expansion physically shatters the rock, producing abundant porosity (Isherwood and Street 1976; Rossi and Graham 2009; Graham et al. 2010).

Heating by forest fires is a physical weathering phenomenon that affects soil minerals directly and can be relatively common in some ecosystems, especially those that produce woody fuels. Burning of fallen logs or large roots can drastically alter soil properties within several centimeters of the heat source. Under severe burn conditions (>500°C), phyllosilicates and gibbsite are decomposed (Ulery et al. 1996), goethite is altered to maghemite (Ketterings et al. 2000), textures become coarser, colors become redder (Ulery and Graham 1993; Ketterings and Bigham 2000), and calcite is formed by carbonation of the white ash produced by thorough combustion of wood (Ulery et al. 1993; Goforth et al. 2005). The spatial extent of these soil mineral alterations depends on fuel loading on the forest floor but often is in the range of 1% of the land area, and has been measured to cover up to 12% (Goforth et al. 2005). The full impact that this weathering mechanism has on soils over the course of thousands of years is unknown.

Chemical weathering is the disintegration of minerals and the rocks they make up, by changes in their chemical composition. A "stability series" proposed by Goldich (1938) illustrates the susceptibility to chemical weathering of some common silicate minerals and generally coincides with empirical observations on stability. The Goldich sequence in order of increasing stability from top to bottom is shown in Figure 4.1. Those geochemists and geologists familiar with the "Bowen's reaction series" of rocks at higher temperatures will recognize this as the converse of that ranking. That is, the least stable minerals are those that crystallize from a "melt" at the highest temperatures. This greater instability is related to their greater disequilibrium with the earth's surface environment, including the soil (Birkeland 1999; Kittrick 1986; Marshall 1977).

In the mafic branch of this series (left side of the "Y"), there is increasing silica tetrahedral linkage with increasing stability from top to bottom. That is, the least stable mineral (olivine) is composed of silica tetrahedra that do not share oxygen atoms. Electrostatic bonds between the tetrahedra and the easily hydrolyzable magnesium ions and the oxidizable iron ions hold the structure together. Also, there is a decrease in the content (percentage) of easily hydrolyzable base cations from the least to the most stable minerals.

In the feldspar branch (right side of the "Y"), there is decreasing isomorphous substitution of aluminum for silicon and decreasing distortion of the lattice from calcic to the potassic feldspars. The bivalent calcium does not fit well into the feldspar structure, though it does satisfy the charge imbalance from considerable substitution of aluminum for silicon. The large monovalent potassium ion is well suited to satisfy the smaller charge imbalance from lesser aluminum-for-silicon proxying and is bonded more energetically in the feldspar structure. Thus, potassium feldspar (e.g., orthoclase) is more stable than the calcium/sodium feldspars (i.e., plagioclase). In quartz, the most stable of the minerals shown, there is complete silica tetrahedral linkage. All oxygen atoms are shared between two silicon ions in ionic–covalent bonds.

The position of biotite in the original weathering sequence should probably be reevaluated. X-ray diffraction, not used to make the original ranking, has shown that biotite alters to vermiculite quite readily. If this simple transformation, which mainly involves the loss of interlayer K, is considered, biotite should move up to at least before hornblende (Figure 4.1) (Nettleton et al. 1970).

As minerals composing bedrock or rock fragments within the solum are chemically weathered, the rock porosity increases so that it has a significant water-holding capacity (Graham et al. 1997; Dreise et al. 2001; Ugolini et al. 1996). These weathered rock materials can then store and supply water (Hubbert et al. 2001a, 2001b) and nutrient cations (Ugolini 2001) for plant uptake. Thus, in many respects, chemically weathered rock material may function as soil in ecosystems (Graham et al. 2010) and is easily disrupted by bioturbations to produce soil material (Graham 2002).

The common general chemical weathering processes of soils, saprolite, and rocks include oxidation, reduction, oxidation-reduction, hydrolysis, hydration, solution, and chelation. These are discussed below.

Oxidation. This is an important weathering process in soil and rock materials in which oxygen supply is high and biological oxygen demand is less than the supply. Basically, oxidation is the chemical process by which an element loses electrons, as in the very common and important oxidation of iron:

$$Fe^{2+} \rightarrow Fe^{3+} + e^-$$

where e^- = electron transfer

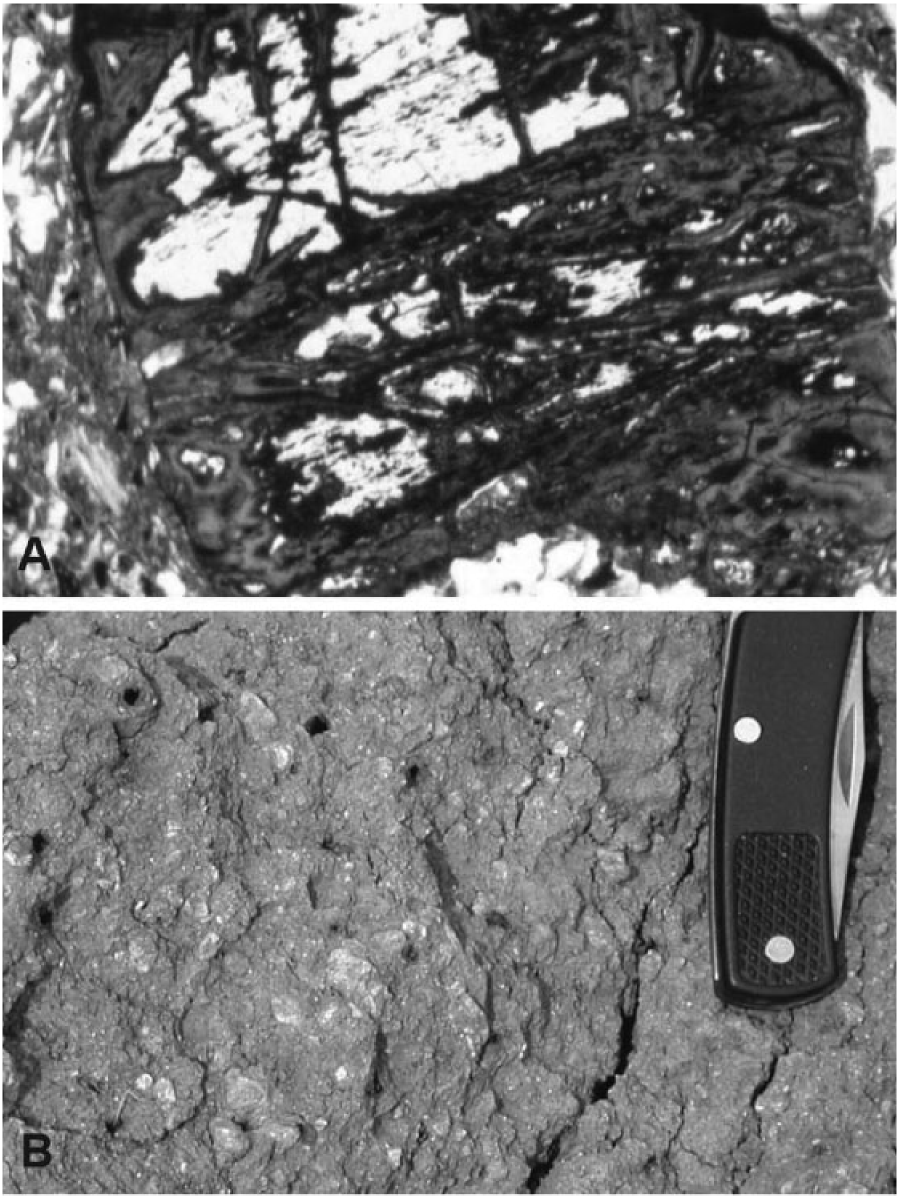

Figure 4.2. (A) Photomicrograph of an Fe-rich garnet (almandine) weathered to produce iron oxides (goethite and hematite). The yellow-brown iron oxides are concentrated in etch pit networks throughout the remnant grain, which appears white in the transmitted plane polarized light. (B) Photograph showing orange mottles of lepidocrocite produced by oxidation around root channels. The reduced soil matrix exhibits the low chroma (gray) color of its component silicate minerals. For color detail, please see color plate section.

Oxidation of iron is a disintegrative weathering process in those common minerals containing ferrous iron as part of their crystal structure. The oxidation of iron disturbs the balance of negative and positive charges in such a way that other cations leave the crystal structure to maintain neutrality, thus causing collapse or increased vulnerability to decomposition by other weathering processes (Birkeland 1999). This happens in the alteration of biotite to vermiculite, for example (Rebertus et al. 1986; Birkeland 1999; Buss et al. 2008), and is important in the weathering of the iron-bearing species of olivine, hornblende, and pyroxene and in glauconite (greensand) (Cloos et al. 1961). Some of the iron released by this disintegration of primary minerals unites with hydroxyl and/or oxygen to form iron oxyhydroxides (e.g., ferrihydrite, goethite, and hematite; Figure 4.2A) (Churchman 2000). Manganese is likewise oxidized and released from primary minerals.

The extent and rate of soil reddening appear to be dependent upon the rate of release of iron by weathering (Schwertmann and Taylor 1989), on the oxygen content of water passing through or standing in soils and underlying materials, and on the species of iron-containing secondary minerals. In some situations there may be oxidation evidence (bright colors) in water-saturated soils and underlying materials where one would expect grayish reduced-iron colors. This may result from well-oxygenated water and/or a low biological oxygen demand caused by the absence of organic matter. In other situations where soil water is well oxygenated, soil color may not reflect oxidation because soils are young and there has been little release of iron by weathering (Birkeland 1999). The landscape position occupied by soils plays a major role, depending on the levels of water saturation and its oxygen content. The red-edge effect (described by Daniels and Gamble 1967) is an example in which soils at edges of Coastal Plain depressions are distinctly redder than their associated soils due to better oxidizing conditions.

The free iron (dithionite extractable) to total iron ratio of soil B horizons can be a good indicator of relative soil development in weakly developed soils (Rebertus and Buol 1985a), directly reflecting the extent of iron oxidation from primary minerals, but also correlating with the amount of clay formed and the extent of clay illuviation. These iron relationships result both from large amounts of oxygenated water moving through the soil, hastening iron release by weathering of primary minerals, and from an absence of reducing conditions.

Reduction. Reduction (gain of electrons) in the geochemical environment occurs where the material is water saturated (such as below the water-table level), oxygen supply is low, and biological oxygen demand is high. The effect is to reduce the iron to the highly mobile ferrous form (Chadwick and Graham 2000). This process is sometimes referred to as *gleization* (Table 5.1). In this form, iron may be lost from the system if there is net downward and outward movement of the groundwater. If the ferrous iron persists in the system, it reacts to form sulfides and related compounds (Fanning and Fanning 1989). The characteristic green and blue–green colors of many reduced soil materials are from either Fe-bearing clay minerals in which the Fe is partially reduced or to green rust minerals (Schwertmann 1993). Green rusts consist of positively charged [Fe^{2+}, Fe^{3+}]–OH layers with various anions (Cl, SO_4, CO_3) between the layers to balance the charge (Schwertmann 1993; Trolard et al. 1997). Green rusts oxidize quickly (<1 hour) upon exposure to the air, but the greenish color arising from clay minerals persists longer. Localized accumulations of oxidized iron in an otherwise reduced matrix often takes the form of lepidocrocite (γ-FeOOH), imparting a characteristic bright orange mottling (Figure 4.2B).

Reduction in soil requires that the oxygen dissolved in the soil solution be removed by microbial respiration. The microbes require soluble carbon for respiration. Total organic carbon content is not a reliable measure of soluble carbon. Dissolved organic carbon (DOC) contents have been observed to increase substantially in late summer in organic-rich soils in Scotland. This increase in DOC was highly correlated with increased iron content in the soil solution (Grieve 1990). The concentration of dissolved organic carbon in soil water of B and C horizons has been found to be approximately 1–2% of that in Oa horizons of Dystrochrepts and Hapludults in the mountains of North Carolina (Qualls and Haines 1991). Addition of soluble carbon from another source creates reducing conditions when inherent soil organic carbon is rather insoluble (Couto et al. 1985).

To reach conditions that reduce Fe^{3+}, first the O_2 in solution must be depleted. Oxygen is utilized as the electron acceptor (i.e., it is reduced); the organic substrate, consisting of reduced carbon, is the electron donor and is oxidized (producing carbon dioxide) in the respiration process. The electron transfer drives metabolic reactions that synthesize high-energy compounds within the microorganisms (e.g., adenosine triphosphate [ATP]). After O_2 is used up, microbes then reduce nitrate to nitrogen gases (mostly N_2 and N_2O) in the denitrification process. Following this, Mn^{3+} is reduced and then Fe^{3+} (Vepraskas 1994). Complete reduction of the iron leaves the soil with a gray or low chroma (chroma of 2 or less) color of the silicate minerals such as quartz and the various clay minerals (Figure 4.2B).

Oxidation-Reduction. Soils can form in originally reduced, subaqueous sediments when they are drained and oxidized. This can come about in response to lowering of regional or local water tables by geomorphic processes, climate change, or land management practices. Alternatively, oxidized soils can become water-saturated so that reducing conditions prevail. These changes are induced when water tables rise in response to climatic or geomorphic processes, beaver activity, or land management practices (e.g., rice paddy culture). A more common process in soil and underlying layers is a seasonal fluctuation from reducing to oxidizing conditions as the water table moves up and down in response to the dynamic water exchange process as controlled by the weather events through the year. The depth to the fluctuating water table is locally controlled by the stratigraphy and geomorphology of the site (Daniels et al. 1987). Iron and manganese oxyhdroxides are commonly concentrated in soils on landscape positions where a fluctuating water table causes alternating oxidizing and reducing conditions (McDaniel et al. 1992; Lee et al. 2001). Fluctuating oxidation and reduction can also occur above a relatively impermeable layer in the soil (an aquitard) that causes a temporary perched water table (McDaniel et al. 2001, 2008).

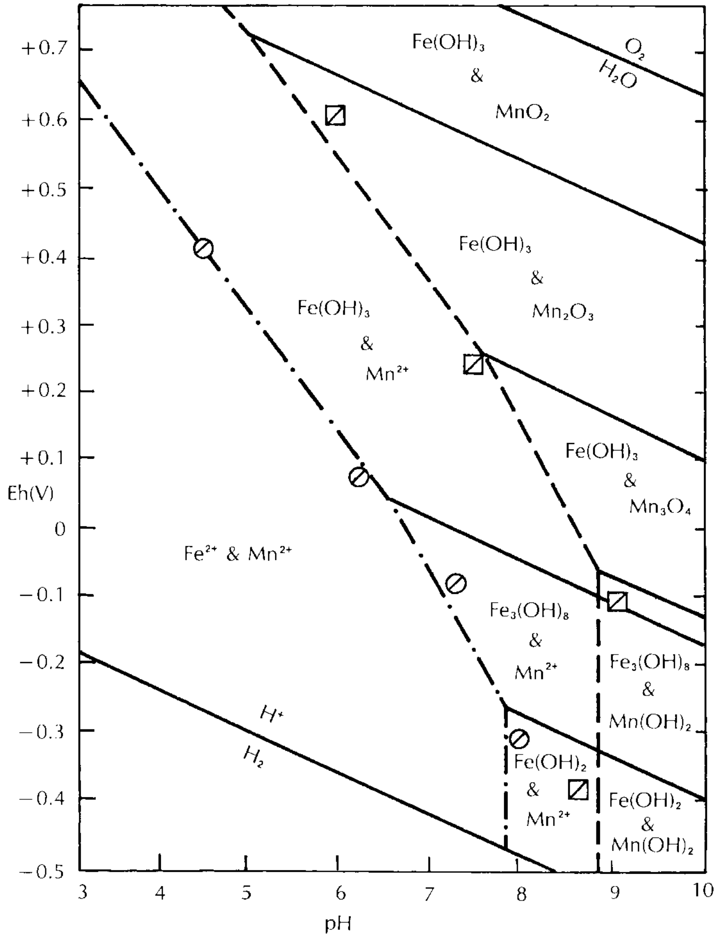

Figure 4.3. Stability fields of iron and magnanese related to Eh and pH in 0.005 M chloride solutions. (After Collins and Buol 1970a)

To understand the effects of fluctuations between oxidizing and reducing conditions on mineral forms, it is useful to refer to phase diagrams (Collins and Buol 1970a, 1970b; Garrels and Christ 1965). In using the example in Figure 4.3, one looks at the pH and the Eh (redox potential) to determine the form of iron and manganese under a variety of conditions. This type of diagram serves as a very helpful geochemical model. If the Eh is not known, one can approximate the intensity of these redox conditions. Some generalizations can be drawn from this chart to illustrate its utility. One such generalization is that with increasing acidity, ferrous iron becomes increasingly stable in more oxidizing conditions. That is, in very acid soil materials, we may expect to find ferrous iron, even though the system is slightly to moderately oxidized. Manganese follows a pattern similar to that of iron but is in reduced form in higher redox conditions at a given pH than is iron. Manganese remains in reduced form at higher pH values than iron. This model explains the deeper occurrences of manganese than iron in the C horizons of soils that have a redox gradient (become more oxidizing with depth) and the redistribution of manganese, but not iron, under mildly reducing conditions produced by short-term soil saturation.

Alternation between reducing and oxidizing conditions is responsible for the release of iron and manganese from primary minerals and their translocation and localization into redoximorphic features within the soil. Iron and manganese are depleted at microsites that when saturated are also well supplied with soluble carbon to energize microbial respiration (Zausig et al. 1993). Walls of root channels containing decaying root tissue are favored sites. Soluble Fe^{2+} and Mn^{2+} are removed with subsequent water flow, leaving a gray redox depletion feature. If clay particles in that area were stabilized by iron oxides, they may disperse and move in the soil solution, creating a clay-depleted micro-site.

Redox concentrations of iron often develop as linings on root channels or other macropores (Plate 4.2B). This occurs when a reduced soil layer first becomes unsaturated and air enters the macropores. If soil adjacent to the macropore remains reduced, soluble Fe^{2+} and Mn^{2+} will diffuse to the aerated macropore and precipitate on the pore wall. Redox concentrations as pore linings are also formed around the roots of rice and other plants that can transport oxygen to their roots when growing in reduced soil (Vepraskas 1994).

Of particular interest and importance relative to silicate clay destruction in soils is the alternation between strong reducing and oxidizing conditions in poorly drained soils, especially those of coastal lowland areas, and in soils with abrupt textural changes at the A–B horizon contact. This process, described by Cate and Sukhai (1964) and Patrick and Wyatt (1964), consists of replacement of exchangeable Al^{3+}, Ca^{2+}, Mg^{2+}, and Na^{+} by exchangeable Fe^{2+} upon onset of reducing conditions. The base cations are especially subject to leaching. With the return of oxidizing conditions, this exchangeable ferrous iron oxidizes to Fe^{3+}. Hydrolysis and precipitation of Fe-oxyhydroxides produces H^{+}. The protons are electrostatically attracted to the octahedral sheet in the clay lattice, reducing structural negative charge there. The charge reduction causes loss of Al^{3+} from the octahedral sheet of the clay lattice, with subsequent disintegration of part of the lattice. Hydrolysis of this Al^{3+} results in additional H^{+} ions, which cause further weathering of the clay (Coleman 1962; Coleman et al. 1960). This process is apparently responsible for the decomposition of 2:1 clays and soil acidity, and may contribute to the formation of localized, clay-depleted microsites and the abrupt textural change itself. These reactions have been termed ferrolysis and are discussed in detail by Brinkman (1970) and van Breemen and Brinkman (1976).

Hydration. Hydration refers to the association of water molecules, or hydroxyl groups, with minerals, often without actual decomposition or modification of the mineral itself. Hydration occurs primarily on surfaces and edges of mineral grains, but in simple salts it may pervade the entire structure, with some changes in properties. An example of the latter case is hydration of the mineral anhydrite to form gypsum:

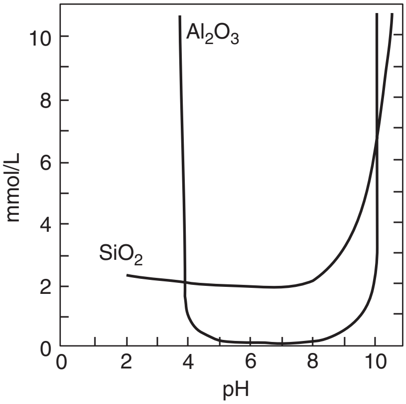

Figure 4.4. Graph showing the relative solubilities of silica and alumina. Silica is more soluble than alumina under pH conditions prevalent in most soils.

$$CaSO_4 + 2H_2O \rightarrow CaSO_4 \times 2H_2O$$

More common are the sorption of water molecules on mineral surfaces and the association of hydroxyls and water in the coordination sphere of aluminum and silica at the broken edges of minerals such as the layer silicates (micas, for example). This sorbed water provides a bridge or entryway for hydronium (hydrated hydrogen) ions to attack the structure. The association of water or hydroxyls with aluminum and silicon at broken edges is but the first step in hydrolysis.

Hydrolysis. This process involves the attack of silicates by the small, highly charged hydrogen ion (which together with its hydration shell is referred to as hydronium), which dissociates from organic and inorganic acids in the soil solution and impinges on minerals of soils and rocks. Equations illustrating this process are presented by Stumm and Morgan (1981) for common aluminosilicate minerals in soils. An example reaction for albite (sodium feldspar), a common soil mineral, follows:

$$\begin{aligned} 2NaAlSi_3O_8 \text{ (albite)} + 2H_2CO_3 + 9H_2O &\rightarrow 2Na^+ + 2HCO_3^- \\ &+ 4H_4SiO_4 + Al_2Si_2O_5(OH)_4 \text{ (kaolinite)} \end{aligned}$$

The H_2CO_3 is the result of CO_2 dissolved in the soil solution.

Note that the products of this reaction include a clay mineral (kaolinite), as well as a base cation (Na^+) and silicic acid in solution. Silica is more soluble than alumina under most soil pH conditions (Figure 4.4), so aluminum is conserved relative to silicon. All of the aluminum and some of the silicon precipitate as kaolinite, while some of the silicon remains in solution and can be removed by leaching, along with the sodium.

In general, hydrolysis reactions with aluminosilicates disintegrate or drastically modify the primary mineral (Casey et al. 1993). New clay minerals are formed by transformation processes in which original structural components are incorporated directly into the clay structure (Banfield and Barker 1994) or by neoformation from

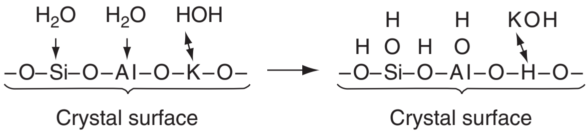

Figure 4.5. Schematic presentation of orthoclase surface reacting with water. Hydration of oxygen ions not shown. (Jenny 1950, © Wiley and Sons, with permission)

solubilized weathering products, including resilication of the aluminum-rich clays formed deep in saprolite layers (Calvert et al. 1980a).

If hydrolysis proceeds rapidly and the soluble weathering products are removed rapidly by percolating water, gibbsite $Al(OH)_3$ can be formed. This situation appears to be quite common in udic and perudic soil moisture regions. More commonly, at least some of the silicic acid released by weathering is short-lived and precipitates with alumina to form short-range-order aluminosilicates (e.g., allophane and imogolite), kaolin, or smectite, depending on solution activities of silica, alumina, and base cations. These newly formed (neoformed) compounds may form coatings or crusts on surfaces of the weathering minerals and thus slow the weathering process (Birkeland 1999). The cations released by hydrolysis can become exchangeable cations attracted to clay mineral surfaces, they may be incorporated in the neoformed products (e.g., magnesium in the octahedral sheet of smectites), or they may be carried out of the weathering environment along with HCO_3^-, after which they may appear in drainage waters or in the groundwater.

The crystal chemistry of the hydrolysis process has been described by Jenny (1950) (Figure 4.5) to be a result of unsatisfied bonds between the oxides and hydroxides of the crystal surface and the cations. Water dipoles are attracted by silicon and aluminum in the crystal to the extent that the water molecules dissociate, enabling the hydrogen ions to combine with oxygen of the crystal surfaces; the hydroxides combine with either silicon or aluminum from the crystal. The hydrogen ions from the water may also displace cations from the crystal lattice, disrupting the crystal surface because of the small size and relatively high charge of hydrogen. Silica tetrahedra and aluminum polyhedra are, therefore, no longer tightly held by the crystal lattice, and they move into the soil solution.

Solution. Solution weathering is the dissolving of calcium carbonate or of simple salts, such as chlorides. The reaction for the carbonate dissolution follows:

$$CaCO_3 + 2H^+ \rightarrow H_2CO_3 + Ca^{2+}$$

This dissolving of calcium carbonate is very important in areas underlain by limestone. The dissolution and removal of the calcium carbonate leaves clay, iron, quartz crystals, and other “impurities” of the limestone as the material from which soils form. The nature of the soil in areas underlain by limestone may be highly dependent upon the nature of the residue left from the dissolution of the calcium carbonate, although in some limestone terrain, eolian-deposited material is an important component of the soil (Delgado et al. 2003). When calcium carbonate concretions and nodules in loess or glacial till parent materials are dissolved and depleted, weathering and other soil developmental processes accelerate. This is because the calcium carbonate maintains a relatively high pH, thereby slowing acid hydrolysis reactions and maintaining high Ca activity in solution, which retards dissolution of other Ca-bearing primary minerals.

Chelation. *Chelate* is defined (Lehman 1963) as "the formation of more than one bond between the metal ion and a molecule of the complexing agent, resulting in the formation of a ring structure incorporating the metal ion." It is important in the stripping of metal ions from some of the primary minerals and in the translocation of these metals in soils (Huang 1989). In soils, the complexing agent, or ligand, is often an organic ion. These chelates originate from certain plant materials or are microbial by-products. The resulting organic molecule-metal cation complex may be stable and persist over a wider range of acidity than otherwise would be the case. The chelates of particular interest to us are organic complexes with aluminum, iron, and calcium. In soils where chelation is prevalent, mineral weathering may be rapid because no inhibitory precipitate forms on the mineral surface, as may happen in hydrolysis. The subsequent removal of the metal ion from a mineral in a rock or in the soil may enhance the downward translocation of the metal in the soil profiles (e.g., in Spodosols, Chapter 17). In other cases, the metal ions stabilize the organic ligand by inhibiting microbial decomposition, thus resulting in an accumulation of the organic–metal complexes in surface horizons.

Potassium Removal from Micas. This weathering process is particularly important in soils in which there is abundant acidity from biological sources and a plentiful source of clay micas originating from the initial material. Removal of a small to moderate amount of the potassium from the interlayers of the mica does not cause great distortion or loss of alignment of the 2:1 structural units (Fanning et al. 1989). Cation exchange capacity is somewhat increased as isolated layers or edges of the mica lose the interlayer K and expand. But with removal of more than approximately 50% of the interlayer potassium, sheet alignment is lost, and strain and distortion of the lattice takes place. Therefore, added potassium is not easily reincorporated in such a way that it is fixed. The mica layers remain open to contact with the soil solution and the remaining interlayer potassium becomes more available. With complete removal of potassium from interlayer planes and replacement by hydrated exchangeable cations, vermiculite and smectite clay mineral types are produced (Fanning et al. 1989).

Products of Weathering

Inorganic materials in soils can be categorized according to origin as inherited, transformed, or neoformed. *Inherited* minerals are supplied by the parent material and are altered only in that they are often comminuted (broken down into smaller

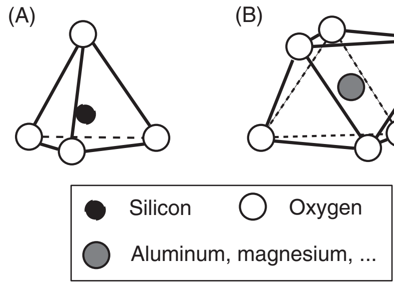

Figure 4.6. The (A) silica tetrahedron and the (B) octahedron are basic structural components of aluminosilicates in soils.

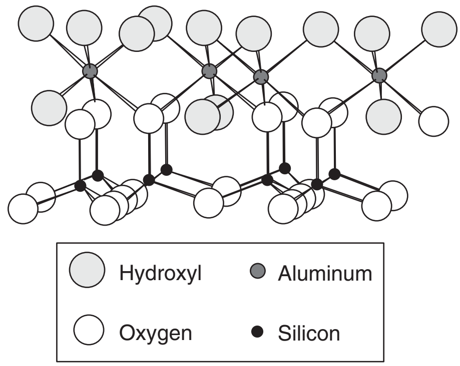

Figure 4.7. Diagram showing how sheets of tetrahedra and octahedra are joined to form the "1:1" clay mineral, kaolinite.

sizes) by physical processes. *Transformed* minerals are those that retain the basic mineral structure of the parent mineral but have undergone minor but important chemical changes. *Neoformed* minerals are those that have precipitated from solution, sometimes on other mineral surfaces that act as templates.

Products of weathering can also be categorized by composition, for example, clay minerals, amorphous aluminosilica compounds, and oxides.

Clay Minerals. Clay minerals are exceptionally important in most soils. They are phyllosilicates, that is, their structures consist of repeating layers. Each layer is composed of two types of sheets. The tetrahedral sheet is composed of silica tetrahedra (Figure 4.6A) that share all of their basal oxygen atoms. In the octahedral sheet, individual ocatahedra are formed by hydroxyls in octahedral coordination around a central Al, Mg, or Fe cation (Figure 4.6B). The individual octahedra are linked into a sheet through the sharing of hydroxyls. The tetrahedral sheet and the octahedral sheet are joined by the substitution of the apical oxygen of the silica tetrahedra for a hydroxyl in the octahedral sheet (Figure 4.7).

If a single tetrahedral sheet is combined with a single octahedral sheet, the clay mineral is referred to as a "1:1 mineral" (Figure 4.7). If the octahedral sheet is sandwiched between two tetrahedral sheets, it is referred to as a "2:1 mineral"

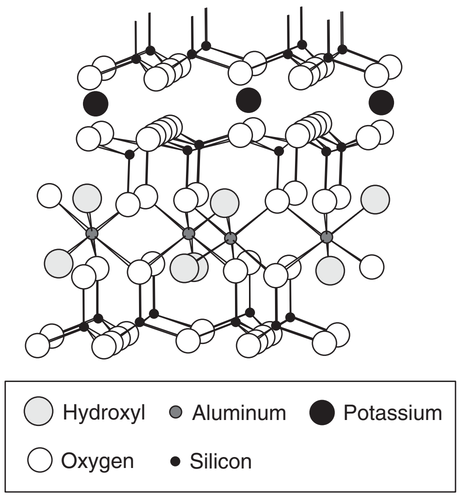

Figure 4.8. Diagram showing how sheets of tetrahedra and octahedra are joined to form a "2:1" mineral, mica. Note that the upper tetrahedral sheet is inverted relative to the lower one and that the apical oxygens of each are shared with the octahedral sheet. Part of the tetrahedral sheet of an adjoining mica layer is shown above the 2:1 structure, and K ions are shown in the interlayer. The K ions bind the 2:1 layers together. Every third cation position in the octahedral sheet is vacant (only two out of three are filled), making this a "dioctahedral" mica—muscovite.

(Figure 4.8). Another important structural feature is determined by the composition of the octahedral sheet. The negative charge from the hydroxyl and oxygen anions is generally satisfied by Al^{3+}, Mg^{3+}, or Fe^{3+}. Other similarly charged cations do not fit so well into the required octahedral coordination. Because two Al^{3+} ions provide the same amount of positive charge as three Mg^{2+} or Fe^{2+} ions, only two-thirds of the octahedral cation sites are filled by Al, but all sites are filled when Mg or Fe are the octahedral cations. As a result, if the octahedral sheet is populated by Al, it is termed "dioctahedral" (two out of three octahedral sites filled), and if it is populated by Mg^{2+} or Fe^{2+} it is termed "trioctahedral" (three out of three octahedral sites filled).

Some phyllosilicates have an inherent, permanent negative charge that arises from isomorphous substitution of a lower charge cation into the structure. For example, when Al^{3+} substitutes for Si^{4+} in the tetrahedral sheet, the four negative charges that had been balanced by the Si^{4+} ion are now only countered by three positive charges from the Al, leaving a net negative charge.

Kaolin is the group name for 1:1 dioctahedral clay minerals including kaolinite (Figure 4.7) and halloysite. Kaolin is a very common weathering product in soils. It is synthesized under conditions of approximately equal concentrations of silicon and aluminum, with high hydronium concentration and essentially an absence of magnesium and other base cations. The formation of kaolin is governed by the relative solubilities of Si, Al, and base cations. Base cations and Si are more soluble than Al under most soil conditions, so they are preferentially leached from well-drained soils. If we consider the weathering of sodium feldspar ($NaAlSi_3O_8$) as an example, the Na and Si are preferentially leached, Si and Al assume equal proportions, and kaolinite ($Al_2Si_2O_5(OH)_4$) may form.

Halloysite is the hydrated, less-ordered species of kaolin. It contains layers of water in its interlayers and often has a tubular morphology (Allen and Fanning 1983). Hallosite crystalizes from amorphous aluminosilica gels (allophane and imogolite), the initial weathering products of volcanic ash. Halloysite also can be a direct product of feldspar weathering (Eswaran and Bin 1978b, 1978c; Southard and Southard 1987) under conditions of high weathering rates, humid conditions, and high leaching rates.

Kaolinite, the other pedogenic kaolin species, is often the most common clay mineral in acid, intensively weathered soils (Dixon 1989). It has a well-ordered stacking of layers, and its formation is aided by the presence of layer silicates as "templates," or patterns, for the 1:1 sheet structure.

Kaolin minerals have little or no isomorphous substitution, so they have virtually no inherent (permanent) charge, and have a very low cation exchange capacity (3–10 $cmol(+)kg^{-1}$).

Mica is commonly inherited in soils as a primary (rock-formed) mineral, but it is rarely formed in soils. Mica has Al substituted for Si in the tetrahedral sheets imparting a high negative charge that is balanced by K^+ between the layers (Figure 4.8). This interlayer K holds the layers together as they stack to give the mineral grain thickness. Although the mica structure has a high negative charge, it is mostly balanced by the interlayer K so that the cation exchange capacity is low to moderate (10–40 $cmol(+)kg^{-1}$). Two main mica species are biotite, which is trioctahedral with Mg^{2+} and Fe^{2+} in the octahedral sheet, and muscovite, which is dioctahedral with Al^{3+} in the octahedral sheet. These structural conditions are very important in determining susceptibility to weathering.

Another species of mica is illite. Although the term *illite* refers to a specific kind of mica (dioctahedral, low charge; Bailey et al. 1984), it has often been used to refer to any clay-sized mica, or even other phyllosilicate species. Unless the species of mica is known for certain, it is best to simply refer to *mica*.

Mica requires moderate to relatively high concentrations of silica and aluminum for stability. Moderate to high acidity causes instability and disappearance, as it is converted to vermiculite, hydroxy-interlayered minerals, or kaolinite.

Chlorite is a primary phyllosilicate that has a structure similar to that of mica except that the interlayer position is occupied by a continuous hydroxide sheet, rather than by K. This hydroxide sheet is similar to the octahedral sheet described previously, except that it is independent from the 2:1 structure (its cations are not directly bonded to oxygen or hydroxyls in the 2:1 structure). Isomorphous substitution gives the hydroxide sheet a positive charge, and it balances the negative charge of the 2:1 structure.

Vermiculite is a 2:1 phyllosilicate often formed by transformation of mica. In the simplest case, all that is required to transform mica into vermiculite is for the interlayer K to be lost. More typically, some of the Fe^{2+} in the octahedral sheet is oxidized and some Fe and Mg are lost from the structure, as well. The loss of K from mica is a diffusion process that can be enhanced by plant uptake of K and the oxidation of structural Fe. Potassium is much more readily lost from biotite than from muscovite, so vermiculite is often abundant in soils derived from granite and another biotite-bearing rocks. Another transformation mechanism by which vermiculite can form is the weathering removal of the interlayer hydroxide sheet of chlorite (Douglas 1989). In any case, vermiculite is formed when soil solutions have moderate pH and high concentrations of silica. The concentration of aluminum in solution must be low, or it will be precipitated in interlayers to form hydroxy-interlayered minerals.

Vermiculite inherits a high structural charge from its precursor mica (or chlorite) and, because the interlayer cations in vermiculite are easily exchangeable, the high structural charge translates into a high cation exchange capacity (100–150 cmol(+)kg^{-1}).

Smectite is a group name for 2:1 phyllosilicates that includes montmorillonite and beidellite, among others. Smectite can neoform when soil solutions have relatively high pH, base cation status, and Si activities. Generally these conditions are met in the vicinity of decomposing mafic minerals and in situations where drainage is impeded by geomorphic position, such as basins, or pedogenic features, such as duripans. Montmorillonite is unstable under conditions of low pH and rapid leaching, but beidellite, an aluminum-rich smectite, is commonly found in the acidic E horizons of some Spodosols (McDaniel et al. 1996).

While smectites often neoform in soils, they are also transformed from vermiculite through loss of structural charge and they may be inherited from parent materials, such as certain shales. Montmorillonite that is inherited from the parent material may persist under conditions of high weathering intensity if it is protected from leaching within dense clay layers (Karathanasis et al. 1986; Borchardt 1989).

Smectites have high cation exchange capacity (80–120 cmol(+)kg^{-1}) and impart a strong shrink-swell behavior to soils in which they are abundant.

Hydroxy-interlayered minerals are formed when the interlayer spaces of vermiculite or smectite serve as a sink for the aluminum in solution (Barnhisel and Bertsch 1989). In either case the result is referred to as a hydroxy interlayered mineral (HIM) because the aluminum precipitates as a hydroxy polymer in the interspace. The hydroxy aluminum may occur with varying degrees of completeness ranging from interlayer islands to nearly complete sheets. Extensively hydroxy interlayered minerals have in the past been referred to as pedogenic chlorite. Aluminum-interlayered minerals are found in largest amounts in surface horizons of acid, well-drained soils, and their concentration decreases with increasing depth in the profile. The cation exchange capacity of the clay is partially blocked and neutralized by this interlayer aluminum precipitation (Jackson 1963a, 1963b; Rebertus et al. 1986). The scavenging of aluminum by the interlayer spaces has an "antigibbsite" effect in that it prevents the precipitation of gibbsite (Jackson 1963a, 1963b).

Allophane and Imogolite. Allophane and imogolite are aluminosilicates with short-range atomic order (Wada 1989; Dahlgren 1994). They are often the first weathering products of volcanic ash and feldspars, especially under humid, leaching conditions (Zelazny and Carlisle 1971). The Al/Si composition of allophane ranges from that of halloysite (Al/Si about 1) to an aluminum-rich composition with an Al/Si ratio of about 2. Imogolite has a composition similar to the aluminum-rich allophane, with an Al/Si ratio of 2, and generally is more crystalline than allophane, as determined by X-ray diffraction studies. Allophane and imogolite are distinguished by their morphologies, based on transmission electron microscopy. Allophane occurs as tiny

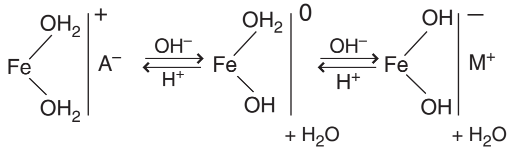

Figure 4.9. Diagram illustrating how the surface charge varies with pH for iron oxyhydroxides. From the point of zero net charge (PZNC), raising the pH of the soil solution (increased OH^- concentration) results in a negative charge on the material, imparting a cation exchange capacity. Lowering the pH (increasing the H^+ concentration) results in a positive charge on the material, imparting an anion exchange capacity. Note that "A^-" denotes an anion and "M^+" indicates a cation.

spheres, whereas imogolite has a threadlike or fine tubular morphology. These two compounds are common in Andisols and in some Spodosols (Farmer et al. 1980; Wang et al. 1986; Dahlgren and Ugolini 1991).

These short-range-order compounds form under conditions of medium to low pH, in moist or wet soil systems, by rapid weathering of volcanic ash or feldspars. Formation is also favored by relatively low soluble carbon contents. Under these conditions, alumina and silica coprecipitate at the expense of aluminum-organic complexes, which tend to form under more acidic conditions (Shoji et al. 1993).

The cation exchange capacity of allophane and imogolite depends on the pH of the soil solution. At a pH of about six, these materials have equal amounts of negative and positive charges on their surfaces (known as the point of zero net charge [PZNC]). At lower solution pH, hydrogen ions associate with the surfaces and impart a positive charge (and an anion exchange capacity), while at higher pH values hydrogen ions dissociate from the surfaces imparting a negative charge (cation exchange capacity). An illustration of pH-dependent charge is presented in Figure 4.9 for an iron oxyhydroxide, and the mechanism is the same for allophane and imogolite.

Oxides and Hyroxides. Soils commonly contain various Al, Fe, and Mn oxides or hydroxides. All of these minerals have pH-dependent cation exchange capacities as described for allophane and imogolite, although the PZNC varies by mineral species.

Gibbsite $Al(OH)_3$ forms and persists under conditions of low silica concentration and low pH, with absence or low concentration of base cations. Gibbsite may form by aging and crystallization of gels, or by desilication of kaolinite or allophane/imogolite (Southard and Southard 1989). It is common in both intensively weathered soils and in soils with actively weathering feldspars (Norfleet and Smith 1989).

Gibbsite is concentrated in soils as a result of at least two seemingly very contrasting processes, but in either case, the stability of gibbsite is determined by the composition of the soil solution, especially the activity of silica. In one case, kaolinite weathers to yield gibbsite upon desilication in soil systems in which there are few other sources of silica. In the other, gibbsite forms by direct weathering of primary minerals if the rate of water movement into and out of the soil is rapid, driving

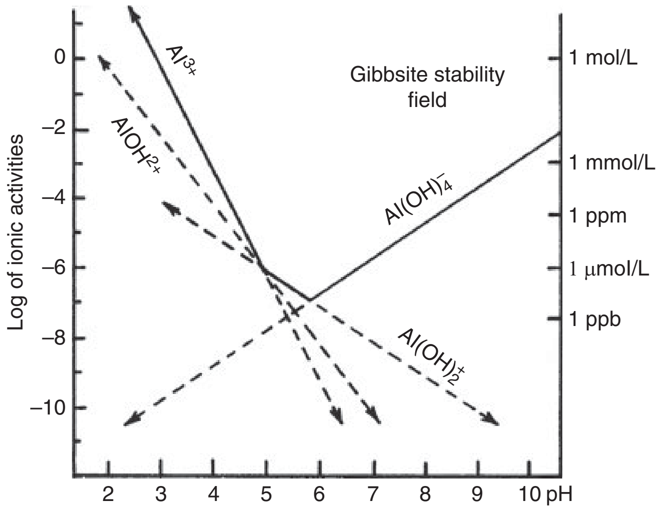

Figure 4.10. Stability field of gibbsite, $Al(OH)_3$. Various Al ions are plotted in relation to pH. (*Left*) Logarithms of activities; (*right*) position of absolute concentrations in millimoles/liter and micromoles/liter. (After Jenny 1980, © Springer-Verlag, with permission)

weathering reactions far toward their endpoint (Graham et al. 1989b). This process is most frequently found at the weathering front very near hard rock, but it may also occur in the solum on steep slopes in udic soil moisture regimes. The rapid removal of silica from the zone of intense weathering leaves an aluminum-rich residue for conversion to gibbsite (Norfleet et al. 1993). Gibbsite is particularly favored to form from the weathering of Ca-rich plagioclase, because it is not only easily weathered (Figure 4.1) but also is rich in aluminum compared to other feldspars.

Gibbsite is most stable in soils of pH range 5–6. Under more acidic conditions, it dissolves, yielding Al^{3+} and 3 OH^- ions in the soil solution and thereby supporting a stock of exchangeable aluminum on permanent charge sites of clay minerals (Jenny 1980). In alkaline soil systems, it dissolves yielding $Al(OH)_4^+$ (Jenny 1980). The stability field diagram for gibbsite is shown in Figure 4.10 (after Jenny 1980). The desilication of kaolinite leaves a gibbsite layer. The equilibrium reaction at pH 7 can be written as follows:

$$Al_2Si_2O_5(OH)_4 + 5H_2O \rightarrow 2Al(OH)_3 + 2Si(OH)_4$$

Kaolinite and gibbsite coexist at equilibrium, in the pH range specified (5–6). With higher $Si(OH)_4$ and Mg concentrations, and at higher pH, montmorillonite is formed by further resilication of kaolinite (Jenny 1980), probably through dissolution of the kaolinite and reprecipitation of the montmorillonite. The stable minerals that form at varying amounts of silica in solution are depicted by Jenny (1980) as follows, based on data of Kittrick (1977):

ppm SiO_2 in Solution				
0	10	50	100	150
gibbsite	kaolinite	montmorillonite	amorphous	silica
(<1)		(1–100)	(100–150)	(>150)

There have also been reports of an "antigibbsite effect" in which aluminum released in the weathering process is shuttled to interlayer spaces of vermiculite or to formation of allophane or other poorly crystalline minerals, leaving little Al for gibbsite formation (Jackson 1963b).

Green rusts are a group of Fe^{2+}–Fe^{3+} hydroxides with variable compositions and structures. A positive charge from the substitution of Fe^{3+} for Fe^{2+} is balanced by anions such as Cl^-, SO_4^{2-}, and CO_3^{2-} (Kämpf et al. 2000). Green rusts occur in reduced, weakly alkaline to weakly acid soils. These compounds account for the greenish blue colors that rapidly change to yellowish brown when water-saturated, reduced soils are brought into contact with the atmosphere. Green rusts are intermediate phases in the abiotic formation of crystalline Fe oxyhydroxides, such as lepidocrocite and goethite.

Ferrihydrite ($\sim Fe_5HO_8 \times 4H_2O$) is a reddish brown, poorly crystalline precipitate. Its formation is favored by rapid release and oxidation of Fe in the presence of high concentrations of organic matter or Si. These compounds inhibit the formation of other Fe oxyhdroxides (Kämpf et al. 2000). Warm, dry conditions favor the rapid transformation of ferrihydrite to hematite. Consequently, ferrihydrite is commonly found as an initial weathering product of mafic minerals in cool, moist soils. It is common in spodic horizons, where microbial decomposition of organometal complexes releases Fe that precipitates as ferrihydrite. It also forms where Fe^{2+}-laden water is exposed to the atmosphere, such as in soils around springs or in macropores (e.g., root channels) of seasonally saturated soils.

Goethite (α-FeOOH) is the most common soil iron oxide or oxyhydroxide, imparting a characteristic yellow brown color (Plate 14.1). It is present in most oxidized soils, even in red soils, where its color is masked by the pigmentation of hematite. Goethite forms by precipitation from solutions with relatively low Fe^{3+} activity. In soils, this often results from the slow release of Fe by mineral weathering, which may be due to low Fe content, inherent resistance to weathering, or environmental conditions (e.g., cool temperatures) that impede weathering. Slow decomposition of organometal complexes in spodic horizons also favors goethite formation. Formation of goethite from ferrihydrite requires dissolution and reprecipitation (Kämpf et al. 2000). Although the iron is in the ferric form, goethite occurs in soils that may experience some short-term saturation.

Hematite (α-Fe_20_3) imparts a red color to soils (Plate 3.1) and is the second most common soil iron oxide. It forms through the dehydration and recrystallization of ferrihydrite under conditions of high Eh and moderate to high pH (Schwertmann and Taylor 1989). Hematite is common in soils derived from parent materials with easily weathered mafic minerals (Figure 3.4). Its abundance, and the red color it imparts, increases in older soils and in soils of warmer climates. Aluminum substitution is common in soil iron oxides and it increases their resistance to reduction and dissolution. Hematite is typically less Al-substituted than goethite, so it is preferentially dissolved under mild reducing conditions (Macedo and Bryant 1989; Bryant and Macedo 1990; Peterschmitt et al. 1996). As a result, redder soils usually occupy slightly better drained parts of the landscape.

Lepidocrocite (γ-FeOOH) is bright orange. It has the same chemical composition as goethite but has a different crystal structure. It is often seen as mottles in noncalcareous soil horizons that are periodically water-saturated and reduced. In reduced soils, ferrous iron can migrate to locally oxidized zones, where ferric iron precipitates as lepidocrocite (Plate 4.1B).

Maghemite (γ-Fe_2O_3) is dark reddish brown. It has the same chemical composition as hematite but a different crystal structure. It forms through the oxidation of lithogenic magnetite, or by heating goethite to about 400°C in the presence of organic matter (Kämpf et al. 2000). The latter mechanism is accomplished in the natural environment through wildfires (Goforth et al. 2005). Maghemite is most common in soils of the tropics and subtropics.

Manganese oxides comprise a variety of minerals and poorly crystalline compounds. They are present in very low concentrations (<0.05%) in most soils. Manganese oxides are very effective black pigmenting agents, and concentrations of even a few percent impart a black color. Manganese oxides are relatively dynamic in response to redox conditions in that they dissolve and precipitate at higher Eh values than do iron oxides (Figure 4.3). Their formation in soils is thus largely controlled by microbial redox conditions, and they are concentrated, often as nodules, in soil horizons that host fluctuating water tables.

Silica. The term *silica* includes those minerals and amorphous forms composed predominantly of SiO_2. The most common of these is quartz, which is pure, highly crystalline SiO_2. Because each O ion in every SiO_4 tetrahedron is shared with a neighboring tetrathedron, quartz has a very stable, bonded structure (Monger and Kelley 2002). Consequently, quartz is very resistant to weathering, as reflected by its place in the Goldich weathering sequence (Figure 4.1). Quartz rarely forms in soils, but it is nearly ubiquitous in them because it is inherited from parent materials and persists as other minerals weather away. Silica that forms in soil is *opal*, a less crystalline and partially hydrated form of silica. This pedogenic silica is typically amorphous (opal-A), but in some very old soils it has intermediate crystallinity (opal-CT).

Silica in crystal (quartz) and gel (opal) forms can be very important in determining the course of clay mineral development in soils because these two forms contribute very different amounts of silicon to the soil solution (Drees et al. 1989). Crystalline quartz dissolves slightly in water to yield 5–20 mg kg^{-1} of SiO_2, whereas amorphous silica precipitated in the lab will maintain 110–150 mg kg^{-1} of SiO_2 in solution at pH 7 (Jenny 1980). The $Si(OH)_4$ formed by hydration of the SiO_4 is adsorbed by soil clay surfaces so that many soils have less than 10 mg kg^{-1} of silica in solution (soils weathered from volcanic ash are a common exception).

The silica freshly released during weathering of volcanic ash or feldspars plays a very active role in the formation of new clay minerals. Where leaching is limited by low rainfall, opaline silica precipitates from solution in dry subsoils and accumulates with age (Kendrick and Graham 2004), eventually producing duripans (Boettinger and Southard 1991; Chadwick et al. 1987a, 1987b; Munk and Southard 1993). Biogenic silica, including phytoliths, diatoms, and sponge spicules, can be an important component of some soils (Meunier et al. 1999). Poorly ordered silica and aluminosilicates have been implicated as temporary cementing agents in hard-setting E horizons and fragipans (Chartres et al. 1990; Karathanasis 1987a, 1987b).

Smeck et al. (1983) pointed out the importance of *kinetics* (reaction rates) in silica solubility in soils. They state that theoretically, the concentration of soluble silicon in soils should be 3 mg L^{-1}, or the solubility of quartz. But soluble silicon levels in soils are actually 15–20 mg L^{-1} (Wilding et al. 1977), and this soluble silica is probably in equilibrium with the silicate coatings on mineral grains. Due to the very slow precipitation rate of quartz, soluble silica concentrations are related to the faster adsorption–desorption of silica on amorphous or poorly crystalline Al- and Fe-oxyhydroxide coatings on clay minerals. This illustrates some of the difficulties in trying to work with pure thermodynamic calculations of mineral stability in soils.

Equilibria and Stability Ranges for Clay Minerals

Phase diagrams are used for quantifying relationships between minerals present and the ionic activity of the solution (Garrels and Christ 1965; Kittrick 1977; Marshall 1977). These mineral equilibria plots are often used together with a plot of the concentration of dissolved constituents in waters (e.g., groundwater, stream water, soil solutions) on the same diagram to determine which minerals are in equilibrium with the solutions of interest (Birkeland 1999; Karathanasis 1991). These types of plots generally confirm that kaolinite is the stable end product of the weathering of silicates (Birkeland 1974), but soil water content is dynamic, and testing equilibria in soil solutions is difficult. Also, as pointed out by Araki and Kyuma (1985), total loss of elements from parent rocks, which has been used to estimate degree of weathering and mineral stability, may not be a suitable indexing method because the weatherability of rocks depends heavily on the mineral weathering potential of each mineral in the rock, rather than on bulk rock chemistry. Lithology of the parent rocks is one of the main factors controlling chemical and mineralogical composition of soils (Araki and Kyuma 1985).

Barnhisel and Rich (1967) found that clay mineral formation in a boulder conglomerate of varying lithology differed widely and was dependent upon the chemical composition of the rocks and the ion concentration in the immediate vicinity of the minerals weathering in the boulders. That is, composition of parent rocks has major impact on weathering products until very late in the weathering process.

These findings suggest that simple ratios like free iron to total iron, weathering and lithology indices, and cation exchange capacities (Araki and Kyuma 1985; Rebertus and Buol 1985a) may be of more immediate use in soil genesis–oriented studies of weathering stage and direction in complex soil systems than are stability diagrams and equilibrium studies. An exception would be for prediction of the final endpoints of weathering and where true equilibrium exists.

We should keep in mind, however, the basic principles of mineral synthesis in soils: ionic concentration and ionic equilibria in soil systems, solubility products of the chemical entities involved, Eh–pH of the environment, and the kinetics of the important coupled weathering–synthesis reactions. Principles of ionic concentration and ionic equilibria, as presented by Garrels and Christ (1965) and Marshall (1977), provide a quantitative base and a model but probably are not a direct working tool in our present state of knowledge. Such approaches should be especially useful in those soil-weathering systems controlled by quartz or amorphous silica forms.

Jenny (1980) cited the strong need for a basic and fundamental thermodynamic approach to clay mineral stability (as advocated by Garrels and Christ [1965]). However, he concluded this approach to be handicapped by the uncertainty of attaining equilibrium and by the analytical requirements of this methodology.

Kittrick (1977) discussed some problems with the use of mineral stability diagrams. He reported that reliable thermodynamic data are not available for most minerals found in soils. Kittrick does indicate, however, that very simple mixtures of minerals can be expected to approach predicted equilibria, but he doubts that actual equilibrium is reached among all soil minerals present in the soil solution. Smeck et al. (1983) indicate there are additional problems in this area of study including the difficulties of sampling the microenvironments surrounding individual mineral grains, the inability to handle and analyze more than a small number of mineral components and ion species, and the narrow range of solution compositions within which one must work.

Soil-forming Processes

5

Soils acquire and maintain their characteristics and composition while undergoing simultaneous alteration by an almost infinite number of physical, chemical, and biological reactions. A soil, like a house, is put together or dismantled by specific processes. Some processes are simultaneously active while others function in sequence. Marbut (1935) stated that nobody has ever seen a mature soil form in toto. Yet, we do observe some processes in operation such as the cracking of clayey soils during dry periods and the incorporation of plant debris into soil by earthworms and ants. Resulting soil features may differ not only in kind but also in durability ("pedologic half-life").

All of the naturally occurring elements are present in soil. Their proportions are determined by the composition of the original parent material, including in some cases subsequent aerosol and/or fluvial deposition, as modified by the within-soil reactions. Water is essential for virtually all soil processes and is present in liquid and vapor phases in all soils. In many soils, the solid phase (ice) is also present, at least seasonally. Soils are inhabited by copious numbers and species of microorganisms nourished by organic carbon entrapped in plants and added to the soil as the plants, or parts of plants, die. Soils are also subjected to physical disturbance by a multitude of insects, rodents, and plant roots.

The possible number of interactions among the inorganic and organic components in soils is staggering. Although laboratory experiments can demonstrate that specific reactions and processes can produce specific soil features, the actual course of events within undisturbed soil will probably never be fully known because the cumulative impact of soil-forming processes spans such long periods of time. Some soil features are ephemeral, others enduring. Some relatively permanent features may be produced by a single episodic event and persist for millennia while other soil properties require undetermined times, perhaps millennia, to form.

Some soils form in materials that have been deposited in single events, such as floodplain alluvium, landslide and mudflow deposits, volcanic deposits, moraines, and spoil piles of earth materials disturbed by human activity. Soils also form and evolve as a weathering profile lowers into and disrupts underlying bedrock (Frazier and Graham 2000). Soil development proceeds much more rapidly in deep, loose clastic deposits than when it must follow downward weathering into solid bedrock. In any case, currently observed soil-forming processes combine to produce soils that are

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characteristic of the environmental conditions (Chapter 3) that collectively promote and nurture these processes. In each principal kind of soil, discussed in later chapters, processes of soil formation occur in unique proportions, intensities, and sequences.

Two scientific approaches have been used in the study of soil genesis, termed static pedology and dynamic pedology by Singer et al. (1978). The first proceeds by obtaining data from field observations and laboratory analysis of samples (NRCS 1996) and then inferring from data what processes could have been capable of producing the observed soil properties, whether or not they operated continuously or intermittently over time. The second approach is to monitor processes in situ, using apparatus such as suction-plate lysimeters that extract samples of percolating water in host horizons (Dahlgren and Ugolini 1989) and troughs wrapped around tree trunks to catch stem flow (Gersper and Holowaychuk 1970a, 1970b). Laboratory simulations using leaching columns and other devices are also used to elucidate soil-forming processes. Whereas the dynamic approach measures some of the current processes, these processes may not accurately reflect the impacts of long-term processes and may miss entirely those processes that operated only for limited periods of time during the formation of a given soil. Generally, some combination of these two approaches provides the most useful information about present and past soil-forming processes.

Conceptualizing the Environment of Soil Formation

Soils are open systems subjected to various intensities and durations of energy and substance dynamics illustrated in Figure 1.8 and Figure 1.9. For a soil property to be present, it must be compatible with the existing environment within the soil, either because currently active soil-forming processes are responsible for its formation or because it formed in past environments and is stable enough to persist under current conditions.

Energy Exchange. Soil temperature is determined by radiation from the sun and by conductivity and thermal diffusion of energy from the air above the soil. Vegetative cover is instrumental in modifying energy exchange between the soil and the air. A major control of the extent of these processes is the location of the soil on the earth's surface. Soils near the equator are subjected to nearly uniform day length and little seasonal fluctuation in solar radiation. Soils located poleward of the equator experience increasingly more seasonal variation in the amount of radiation they receive, and thus at higher latitudes, have more drastic seasonal temperature changes.

Except during some seasons in polar latitudes, all soils are subject to daily reversals of energy flow through their surface. During daylight hours, the surface of the soil gains energy and becomes warmer, whereas at night, heat is lost by radiation, and the surface cools. The magnitude of temperature changes, both daily and seasonally, is most extreme in the near-surface parts of the soil and is less extreme with depth.

Geographic location is a major control of average soil temperature and seasonal dynamics. Soils near the equator (tropical latitudes) experience little seasonal temperature change below 50 cm while greater seasonal changes occur at higher latitudes. In near polar locations the temperature at depth never exceeds 0°C and the soil remains frozen (permafrost), although soil near the surface seasonally freezes and thaws.

Daily temperature dynamics in the upper 50 cm are experienced in all soils (Figure 3.6). These daily temperature reversals with depth drive vapor movements in the soil. As the land surface cools in the evening, gasses, including CO_2 and water vapor, are expelled through the soil surface and water condenses as dew on the soil and/or grasses. As the sun rises each day, surface temperatures increase and vapors are driven downward.

External features modify energy exchange reactions. If there is vegetative cover, such as forest that shades the soil, surface temperatures remain cooler during daylight hours and warmer at night as compared with unshaded soil (e.g., Johnson-Maynard et al. 2004). Aspect is a microclimatic factor. Hill slopes facing away from the sun (north-facing aspects in the northern temperate zone) do not heat as much as those aspects facing into the midday sun. Light-colored soil surfaces reflect more sunlight than dark-colored ones. Soils that contain more water, either because of their landscape position or higher clay content, require more energy to raise their temperature and require dissipation of more energy to lower their temperature than do drier soils. Thus, maximum soil temperature of surface and near surface horizons during daylight hours is lower in wet or fine-textured soils than in dry or sandy soils.

Energy, measured as soil temperature, is an intricate component of all the biogeochemical reactions within the soil, affecting both rate and type of reaction.

Water Exchange Processes. Like energy, water is instrumental in all biogeochemical reactions within the soil. Furthermore, both the physical impact of raindrops and the kinetic energy of runoff water have major influence on erosion and deposition. The conditions present at the soil surface and the slope gradient of the soil surface determine to a large extent whether precipitation enters the soil or runs off over the surface of the soil.

The water exchange processes within the soil are dynamic and largely controlled by weather events, but are altered by pore size distributions within the soil. Water enters the soil as a liquid but can leave the soil as a vapor, as direct evaporation to the air or by being sorbed by plant roots and transpired from the stomata of the plant leaves. Water can also leave as a liquid through the lower layers of the soil in the process of leaching. While water is in the soil, it is a solvent for minerals and is an essential ingredient for both macro- and microbiological respiration.

The infiltration of water through the soil surface is a sporadic event. Most water enters only during rainfall events or upon the melting of snow. Downward movement of water attains its greatest velocity in the larger pores, where it is capable of transporting suspended colloidal particles of clay and organic material as well as soluble compounds. After the entry of water into the soil ceases, percolation slows and water is distributed via capillary action into smaller pores, where filtering capability of the soil increases (Kretzschmar et al. 1994; Frazier et al. 2002). After the large pores are emptied, water is retained in the medium and small pores in the soil. During this time, there is some dissolution of soil minerals, and the ionic concentration in the water increases. If no further entry of water takes place and the soil is not frozen, the soil water retained in the medium-sized pores is removed either by evaporation into the larger pores, with some loss to the atmosphere through the soil surface, or by sorption by plant roots or other soil organisms. In these processes, the water may leave behind suspended material and/or solutes that are selectively excluded by the plant root (a filtering or sieving process). When soils become very dry, water may exist only as very thin films on mineral grains or on soil aggregates. As the volume of water is decreased by these processes, precipitation of solutes and suspended material occurs. The depth at which this takes place varies with each rain event, depending upon the amount of water added, the water content of the soil at the time of the rain, and the rate of evapotranspiration. If rainfall events occur before sufficient pores are emptied, soil solutions in the larger pores are forced deeper by the new water. If the water is moved to a depth at which it can no longer be extracted by roots, it is termed leaching water, and it proceeds as "pushed" by subsequent rainfall into the groundwater, often well below rooting depth.

Lateral redistribution of water frequently occurs among soils on sloping land. If the rate of water entering a soil body exceeds the rate at which it can percolate downward through a less permeable horizon in the subsoil, water will flow downslope above the less permeable horizon into an adjacent soil (McDaniel et al. 2008).

Water movement within the soil is the mechanism of transport for almost all mobile soil constituents. The accumulation of any constituent that can be suspended or dissolved in water is related to the repetitive water-exchange process (Table 5.1).

Leaching. Leaching is analogous to eluviation by solution but connotes removal from the entire solum, although it is common to speak of an eluviated E horizon as a leached horizon thereby referring only to leaching from a specific horizon within a soil. Leaching is a primary process, prerequisite in many soils to removal of both colloids and solutions.

When referring to leaching as removal from the entire soil, Polynov (1937) compiled a relative mobility index, with 100 being the most mobile of some soil constituents, as based on analyses of igneous rocks and of dissolved loads of rivers: Cl^-, 100; SO_4^{2+}, 57; Ca^{2+}, 3.00; Na^+, 2.40; Mg^{2+}, 1.30; K^+, 1.25; SiO_2, 0.20; Fe_2O_3, 0.04; Al_2O_3, 0.02. The abundance of these same elements or compounds in the earth's continental crust has been calculated by Wedepohl (1995), and are expressed as percent Cl^-, tr; Ca, 3.85; Na, 2.36; Mg, 2.20; K, 2.14; SiO_2, 61.5; Fe_2O_3, 6.28; Al_2O_3, 15.1. While being of general interest on a global basis, such values are of little interest to soil formation processes wherein total leaching of colloids and solution is a sporadic event. Partial leaching, with sporadic precipitation of Mg, K, Si, and Al in soil clay minerals, and Fe as ferric oxyhydroxides in nodules, concretions, and ironstone deposits within the soil is more germane to the formation of soil horizons.

Table 5.1. Some processes of soil formation that are complexes of subprocesses and reactions

Term	Fourfold Categorization[a]	Brief Definition
1a. Eluviation	3	Movement of material out of a portion of a soil profile as from an albic horizon
1b. Illuviation	3	Movement of material into a portion of a soil profile as in an argillic or spodic soil
2a. Leaching (depletion)	2	General term for washing out, or eluviating, soluble materials from the solum
2b. Enrichment	1	General term for addition of material to a soil body
3a. Erosion, surficial	2	Removal of material from the surface layer of a soil
3b. Cumulization	1	Eolian, hydrologic, and human-made additions of mineral particles to the surface of a solum
4a. Decalcification	3	Reactions that remove calcium carbonate from one or more soil horizons
4b. Calcification	3	Processes including accumulation of calcium carbonate in Bk and possibly other horizons of a soil
5a. Salinization	3	The accumulation of soluble salts such as chlorides, sulfates, and bicarbonates of sodium, calcium, magnesium, and potassium in salic horizons
5b. Desalinization	3	The removal of soluble salts from salic soil horizons
6a. Alkalization	3	Accumulation of sodium carbonate raises pH above 8.5
6b. Dealkalization	3	Leaching of sodium carbonate lowers pH below 8.5
7a. Lessivage	3	The mechanical migration of small mineral particles from the A to the B horizons of a soil, producing B horizons relatively enriched in clay, as in argillic horizons
7b. Pedoturbation	3	Biological, physical (freeze-thaw and wet-dry cycles) churning and cycling of soil materials, thereby homogenizing the solum in varying degrees
8a. Podzolization	3, 4	The chemical migration of aluminum and iron and/or organic matter, resulting in the concentration of silica (i.e., silication) in the layer eluviated
8b. Desilication (ferrallitization, ferritization, allitization)	3, 4	The chemical migration of silica out of the solum, thus the concentration of sesquioxides in the solum (e.g., goethite, gibbsite), with or without formation of ironstone (laterite; hardened plinthite) and concretions
8c. Resilication	4	Formation of kaolinite from gibbsite in presence of excess $Si(OH)_4$ in solution or formation of smectite from kaolinite in presence of large amounts of $Si(OH)_4$ at higher pH values
9a. Decomposition	4	The breakdown of mineral and organic materials
9b. Synthesis	4	The formation of new particles of mineral and organic species
10a. Melanization	1, 3	The darkening of light-colored, unconsolidated, initial materials by admixture of organic matter (as in a dark-colored A horizon)

continues

Table 5.1. *Concluded.*

Term	Fourfold Categorization[a]	Brief Definition
10b. Leucinization	3	The paling of soil horizons by disappearance of dark organic materials either through transformation to light-colored ones or through removal from horizons
11a. Littering	1	The accumulation on the mineral soil surface of organic litter and associated humus to a depth of less than 30 cm
11b. Humification	4	The transformation of raw organic material into humus
11c. Paludization	4	Processes regarded by some workers as geogenic rather than pedogenic, including the accumulation of deep (>30 cm) deposits of organic matter as in mucks and peats (Histosols)
11d. Ripening	4	Chemical, biological, and physical changes in organic soil after air penetrates previously waterlogged material
11e. Mineralization	4	The release of oxide solids through decomposition of organic matter
12a. Braunification, rubification, ferrugation	3, 4	Release of iron from primary minerals and the dispersion of particles of iron oxides or oxyhydroxides in increasing amounts. Soils are colored brownish, reddish brown, and red, respectively
12b. Gleization	3, 4	The reduction of iron under anaerobic soil conditions, with the production of bluish to greenish gray matrix colors, with or without yellowish brown, brown, and black mottles, and ferric and manganiferous concretions
13a. Loosening	4	Increase in volume of voids by activity of plants, animals, and humans and by freeze-thaw or other physical processes and by removal of material by leaching
13b. Hardening	4	Decrease in volume of voids by collapse and compaction and by filling of some voids with fine earth, carbonates, silica, and other materials

[a]The four categories (see text) are (1) additions to a soil body, (2) losses from a soil body, (3) translocation within a soil body, and (4) transformation of material within a soil body.

All leaching involves the physical movement of water downward within the solum. General climatic data are of some value in understanding processes of leaching, but water movement in the soil resulting from individual weather events is more specifically related. (See Water Exchange Section.) At the terminus of downward movement, water is removed by evaporation or in the transpiration stream by uptake through root membranes. Ions and compounds are thus concentrated in the soil solution and, depending upon specie populations, form new solid compounds. Most often these are alumina-silica clay minerals, but iron and aluminum oxides precipitate where silica concentrations are low.

Carbonate leaching is of specific interest in soils formed from limestone, carbonate-cemented sandstone, calcareous loess or glacial drift, and sites receiving deposits of calcareous dust. Carbonates precipitate at the terminus of the downward water flow (Plates 10.1 and 10.2).

Lateral Losses and Gains. Within most landscapes, there is movement of material from one soil to adjacent soils. Movement can take place both on the soil surface and below the surface.

Surficial erosion refers to lateral removal of material from the surface of a soil by raindrop splash, runoff waters, or wind. Landslides and other processes of mass-wasting may move entire soil bodies. Surface erosion is most active on hillslopes and results in transfer of material to lower parts of the slope or the valley bottom. Only a portion of the eroded material is removed in stream flow from the area.

Cumulization (from *cumulus*, to heap) (Table 5.1) refers to the accumulation of mineral material by wind or water on the surface of the soil. Cumulization (also called accretion) may be considered a geogenic rather than a pedogenic process. The effects of this process are evident at the base of hillslopes, where material eroded from soils upslope has accumulated. Termed hillslope sediments (Daniels and Hammer 1992), these thin deposits are on slopes of 1 to 20%. They often lack any easily recognizable features because they slowly accumulate and are easily assimilated into the soil profile.

Erosion and cumulization are commonly recognized as paired processes and are often spatially associated on the landscape, but wind-eroded aerosol dust is deposited to some extent on all soils and can be considered part of a 'long-range' loss/gain system (Simonson 1995). Dust deposits are substantial in many areas of west Africa (Moberg et al. 1991), Hawaii (Stewart et al. 2001), and most of the Great Basin and Mojave Desert regions of the western United States (McFadden et al. 1987; Chadwick and Davis 1990; Reheis et al. 1995; Reynolds et al. 2006). Dust deposition and accumulation under desert pavement are key to the development of vesicular horizons in arid landscapes worldwide (Chapter 10, Figure 10.6).

Lateral downslope movement of water through the soil may occur as saturated or unsaturated flow (Ward 1984; Zaslausky and Sinai 1981). This process, known as throughflow or interflow, can transport solutes and suspended material from one soil downslope to an adjacent soil. In udic soil moisture regimes and acid parent materials, soils on the footslopes are typically more acid and contain more manganese than soils on adjacent upslope positions (McCracken et al. 1989). Throughflow waters appear responsible for the manganese accumulation (McDaniel et al. 1992).

Intrasolum Translocations and Transformations

Soil formation involves a complex suite of processes including both complicated reactions and comparatively simple rearrangements of matter, which intimately affect the host soil. Various processes may take place simultaneously or in sequence that mutually reinforce or contradict one another (Rode 1962; Simonson 1959). For example, calcification and podzolization (Table 5.1) operate concurrently in certain

Udalfs. Generally, sequences of processes occur when the results of one process trigger the initiation of a subsequent process. An example is the decalcification of a soil horizon, which leads to the weathering of other calcium-bearing primary minerals to produce clays that can be dispersed and translocated. In this example, the depletion of calcium carbonate is the "pedologic threshold" that must be crossed before clay translocation begins. The example of long-term weathering of kaolinite to gibbsite (Chapter 4) is another example of a threshold. Gibbsite forms only when the cumulative effects of previous processes produce a soil solution with the appropriate silica, alumina, and cation concentrations. In this context, it is important to emphasize that the actual process of weathering kaolinite to produce gibbsite, in and of itself, is not a slow, long-term process. Instead, the whole sequence of other processes that ultimately produce a soil system wherein gibbsite is stable is, in the aggregate, a slow process.

The results of a given process may tend to maintain the soil in its current condition, or it may tend to change the soil. Some processes, such as crystal growth in a void in the interior of a large boulder and movement of fluids inside a large tree root or in a hibernating rodent in a soil, are remote from the active part of the soil and, hence, even though they take place within the soil, may be excluded from the long list of soil-forming processes. On the other hand, crystal growth at the surface of the buried boulder, exchange of fluids between soil and root hairs, and exchange of gases between dormant rodents and the soil atmosphere are intimate parts of the processes of soil genesis. Pedogenic processes include gains and losses of materials from a soil body in accordance with the degradational, aggradational, or intermediate geomorphic character of the site, as well as translocations within a soil body.

Two overlapping trends in soil development are horizonation and haploidization. Simply stated, there are processes that tend to mix soil, and there are processes that tend to form layers or horizons within soil. Both trends are present in all soils, and the degree to which certain soil properties are present represents the relative intensity of these contrasting forces.

Biocycling. In its broadest sense, biocycling is a twofold movement of chemical elements: (1) elements are removed from soil by the plant wherein organic compounds are formed, and (2) elements are returned to the soil as organic compounds when plants, plant parts, and animals that consume plants die.

Biocycling by higher plants involves numerous biochemical transformations. For the purpose of understanding the major role of biocycling in soil formation, we may focus on a relatively simple sequence of events, as follows. Plant tissue is composed primarily of carbon (45%), oxygen (43%), and hydrogen (6%). Plants obtain carbon and oxygen as carbon dioxide from the air through the stomata of the green aboveground parts of the plants. Water ingested mainly through the roots provides hydrogen and oxygen. The other elements required for growth and metabolism, approximately 6% of a plant's composition, are obtained as inorganic ions in the soil solution around its roots. Within their cells, plants synthesize carbohydrates, fats, proteins, nucleic acids, and other organic compounds. When part or all of the plant dies, the organic compounds become the substrate for soil microorganisms that oxidize them, producing carbon dioxide that is exhausted to the air. As this decomposition proceeds, the essential plant nutrient elements (nitrogen, phosphorus, potassium, sulfur, etc.) are released from the organic compounds into the soil solution as inorganic ions. Microbial decomposition of organic matter is not instantaneous. The microbes of each generation convert about 75% of the carbon they oxidize to CO_2 but retain the remainder in their cells to be oxidized by subsequent generations. The chemical structure of the organic compounds remaining after a series of decomposition stages is less easily oxidized and is known as humus.

Although many identifiable reactions are involved, the net result of biocycling during soil formation is the physical extraction of plant essential elements from the entire volume of soil explored by the plant roots, and their concentration wherever the dead plant material is deposited and decomposes. Of course, organic carbon accumulates in the soil along with the biocycled mineral-derived nutrients. Generally, a greater proportion of dead biomass is deposited on the soil surface as litter, especially in forested ecosystems, but decomposing biomass is also produced by roots, insects, and burrowing animals that die below the soil surface.

When the biocycling process continues, and is uninterrupted by the removal of plant biomass from the site, the surface layer of the soil receives a net influx of plant essential elements, whereas the deeper parts of the soil explored by the roots experience a net loss of those elements. Moreover, as the organic compounds are decomposed, inorganic ionic forms of nitrogen, phosphorous, potassium, and other elements are produced that are readily available for uptake by actively growing roots and soil microorganisms. Thus, upper horizons (epipedons) contain greater amounts of readily available plant essential elements than do the subsoil horizons. Epipedons also usually contain more organic carbon than most subsoil horizons because organic compounds are the vehicle by which essential elements extracted from the soil by plants are returned to the soil.

In many ecosystems fire is a recurring phenomenon that instantaneously mineralizes nutrients from plants and organic matter in O horizons. Burning also produces charcoal, which is a common constituent in many A horizons (e.g., MacKenzie et al. 2008).

Horizonation Processes. Processes responsible for the presence of identifiable horizons in soils include (1) additions of more organic or mineral material to one depth range in the soil than to the other parts of the soil, (2) preferential losses of material from a portion of the soil, (3) translocations of materials from one point to another, usually vertically within the soil, and (4) transformation of mineral and organic substances and/or rearrangement of soil material into structural units or peds preferentially within the soil (Simonson 1959).

The terms listed in Table 5.1 are useful in expressing these processes at different levels of generalization. Some terms are rather specific, relating to the movement or deposition of specific soil constituents. Others are general in usage, conveying only the broad context of the process and often including some of the more specific processes named. Many processes indicate two parts of what can be thought of as a continuous process. Insofar as possible, these are paired in Table 5.1 and in the following discussion.

The most general terms to express the transport of soil material or constituents from one horizon to another horizon in which it accumulates are *eluviation* and *illuviation* (words analogous to *emigration* and *immigration*). An eluviated horizon is one where eluviation or removal of a constituent has taken place; an illuvial horizon is one where a constituent has accumulated. Eluviation as a process consists of mobilization reactions and the translocation of the mobilized constituent. Illuviation involves translocation and reactions that immobilize the translocated constituent. Leaching, enrichment, erosion, and cumulization are also general terms, which have been discussed earlier in this chapter.

Decalcification is specifically used for the eluviation of carbonates, if ever present, within a soil body. The process may lead to the complete removal of carbonates from the entire soil, as is common in more humid areas, or may be accompanied by *calcification,* which is the accumulation of carbonates commonly observed in lower horizons of soils in more arid regions (Plates 10.1 and 10.2). The general reaction involved in carbonate movement is as follows:

$$CaCO_3 + H_2O + CO_2 \leftrightarrow Ca^{2+} + (HCO_3^-)_2$$

Decalcification can be thought to occur when H_2O + CO_2 are present and the reaction moves to the right with the formation of the soluble bicarbonate. *Calcification* occurs when either CO_2 or H_2O is removed from the system and the reaction moves to the left. Further discussion of the process of calcium carbonate accumulation in soils is presented in Chapter 10 (Aridisols).

Desalinization is most frequently used with reference to the removal by leaching of soluble salts from horizons or whole soil profiles that have previously contained enough soluble salt to impair plant growth. Therefore, it is a process that can be active only after soluble salts have accumulated, that is, after salinization.

Salinization operates chiefly in subhumid, semiarid, and arid regions, and some coastal humid regions, wherever a part of the soil is enriched in salts faster than they are leached. Solubilities of common compounds, in grams per 100 mL of pure water (at 0°C unless otherwise indicated), are listed by Hodgman et al. (1962): K_2CO_3, 112; $CaCl_2$, 59.5; $MgCl_2$, 54.3 (20°C); NaCl, 35.7; KCl, 27.6; $MgSO_4$, 26.0; Ca (HCO_3), 16.2; $FeSO_4$, 15.7; K_2SO_4, 12.0 (25°C); Na_2SO_4, 4.8; $CaSO_4$, 0.2; $MgCO_3$, 0.01; $CaCO_2$, 0.001 (25°C); FeS, 0.006 (184°C). Salt accumulation occurs on or near the soil surface as water is evaporated. The soil water may have accumulated the salt by desalinization of soil or geologic material as it flowed in the groundwater or as surface runoff into depressional areas of low permeability. Eolian redistribution of salts from playas to upland areas is a common process in arid landscapes (e.g., Reid et al. 1993). Irrigation of crops, even with only very slightly saline water (e.g., electrical conductivity of 0.7 dS/m), is a common cause of salinization if steps are not taken to leach the salts and provide for drainage. Sulfates, chlorides, and bicarbonates are the predominant salts. Nitrates and borates are generally less common, but nitrates are prevalent in soils under desert pavement in the Mojave Desert (Graham et al. 2008).

Alkalization results in soil pH values greater than 8.5. It occurs by two primary mechanisms, both of which involve sodium (Chadwick and Graham 2000). The first is the result of sodium carbonate accumulation, which is relatively common in arid soils. Sodium carbonate is highly soluble and its dissolution yields abundant alkalinity:

$$Na_2CO_3 + H_2O \rightarrow 2Na^+ + OH^- + HCO_3^-$$

A second, less common, mechanism for alkalization is the hydrolysis of Na on cation exchange sites:

$$NaX + H_2O \rightarrow HX + Na^+ + OH^-$$

where X represents clay or organic matter exchange sites.

Another type of alkalization can take place in patches on forest soils when logs or other heavy fuels are combusted to white ash. White ash consists of the oxides of the base cations contained in plant tissue and can produce pH values on the order of 10 to 12, though carbonation reduces the pH to about 8 within months (Ulery et al. 1993; Goforth et al. 2005).

Dealkalization refers to lowering of soil pH below 8.5 through the removal of sodium carbonate or Na^+ ions from exchange sites. This is often accomplished by displacement of the exchangeable Na by Ca and removal of the Na by leaching.

Lessivage, the translocation of fine clay, and lesser amounts of coarse clay and fine silt, in suspension down cracks and other voids in a soil body, is reflected in (1) depletion of A and E horizons of clay, (2) enrichment of clay in the B horizon relative to the C and/or A horizon (Plate 18.1), (3) higher fine clay to total clay ratio in the B horizon than in the A horizon, and (4) presence of argillans in the B and C horizons. The mobile clay involved may be a product of weathering in the A horizon, may have been present in the parent material initially, may be of eolian origin added to the soil during development (Buol and Hole 1961; Khalifa and Buol 1968), or may be formed and translocated within the B horizon itself.

Naturally occurring humic substances promote dispersion and translocation of clay in soils (Jenny and Smith 1935; Durgin and Chaney 1984; Kretzschmar et al. 1993). These humic substances prevent flocculation by complexing cations in solution, thereby keeping the cations from bonding colloid particles to each other, and by specifically adsorbing to positively charged mineral edges, so those edges do not bond to negatively charged mineral surfaces. The mobility of clay particles is also affected by their size and mineralogy (Kaplan et al. 1993). The very smallest particles, such as neoformed smectites and iron oxides, are most readily translocated, even by low velocity water flow that occurs in the microporous soil matrix. Coarser clay particles, and even silt, can be moved by higher velocity flows that are attained by water moving through macropores. Macropore flow events are less frequent, but they can move more clay and larger particle sizes to greater depths than can flow through micropores. Minerals that tend to occur as larger particles, such as kaolin, mica, vermiculite, quartz, and feldspar, require these larger flow velocities to be moved, while smaller clays, especially those with a strong negative charge such as smectites, are more mobile. More discussion of clay translocation and accumulation can be found in Chapter 8 (Alfisols) and Chapter 18 (Ultisols).

Podzolization is defined as the process by which sesquioxides (Fe- and Al-oxyhydroxides) are translocated in a soil profile (Stobbe and Wright 1959). The process can be explained only in part by the solubilities of ferrous and ferric iron. The soluble ferrous iron forms at the sites of eluviation, and the insoluble ferric iron forms at the point of illuviation. Redox reactions cannot account for the distribution of aluminum; thus, other mechanisms must be involved. This process and the role of chelation are discussed further in Chapter 17 (Spodosols).

Oosting (Edelman 1950) reported that in some soils in the Netherlands, the color of the root zone depends on the vegetation: "Under oak trees the soil is dark-brown, under beeches a bit more reddish; under fir trees orange-brown and under birches more yellowish." These observations suggest that different organic compounds have been synthesized at the different sites and that those compounds and other environmental factors may also affect the forms of iron associated with the organic compounds.

Desilication refers to processes that remove silica from the soil. *Resilication* refers to the addition of silica to a clay mineral, thereby converting it to another species. Aspects of silica dynamics in soils are addressed in Chapter 4. Mineral *decomposition* by weathering also is addressed in Chapter 4.

Decomposition of organic matter involves complex reactions among complex organic compounds. A general scheme of organic matter decomposition is shown in Table 5.1. Various aspects of organic matter decomposition are addressed by several more specific process names. *Melanization* and *leucinization* refer to changes in color value in soil, whether caused by addition or losses, respectively, in the content of organic matter (the common case), or by transformations from dark-colored (melanized) to light-colored (leucinized) organic compounds or vice versa. More discussion of organic matter dynamics in soils can be found in Chapter 13 (Histosols) and Chapter 15 (Mollisols).

The word *littering* is offered here for the accumulation of plant and associated faunal debris (together, an Oi horizon) on the mineral soil surface. The further decomposition or *humification* of such debris is often expressed in an Oe or Oa horizon.

Paludization (*also* paludification) is the accumulation of a thicker mass of organic materials in a poorly drained site where preservation under anaerobic conditions has allowed a net gain through time. This process is geogenic in that it is an accumulation of organic material to such a thickness that Histosols are formed. (See Chapter 13.)

Release of mineral components of organic matter through decomposition is the complex process of *mineralization*. The vast stores of plant essential nutrients in the standing biomass of a forest ecosystem and in soil organic matter of grassland ecosystems are unavailable to plants except through mineralization. This process proceeds through the conversion of carbon in organic matter to CO_2 by fire or by the slower oxidation of microbial decomposition.

Under the very special conditions of polder construction, air is able to penetrate previously unaerated, reduced soil material. The initial chemical, physical, and biological reactions that take place are collectively known as the process of *ripening*. The types of reactions and the resulting soil properties are largely dependent upon the nature of the initial material, which can be either organic or mineral (Pons and Van Der Molen 1973).

The general trend of subsoil reddening in uplands along transects from the polar to equatorial regions has been noted (Schwertmann and Taylor 1989). Three processes, *braunification, rubification,* and *ferrugination* (development of brown, reddish brown, and reddish soil colors), are self-explanatory as to soil appearance. Although a number of mechanisms may be involved, the conditions that determine the goethite/hematite ratio are often of critical importance (Chapter 4), as are interactions of iron oxides with soil organic matter (Chapter 17, Spodosols). In certain very young soils, organic matter itself can have a reddening effect (Turk et al. 2008).

Gleization in poorly drained soils involves the reduction of iron and its segregation into redoximorphic features, or its removal by leaching from the gleyed horizon (Chapter 4). Formation of iron sulfide (FeS) is possible in gleyed soils that are saturated with brackish water. Upon drainage, these soils may become acid from formation of H_2SO_4 through oxidation of the iron sulfide.

Loosening and *hardening* are very general terms useful in qualitative expressions relating to physical alterations of a soil.

Development of Soil Structure. Structural development and expression in B horizons refers to a combination of processes and mechanisms that, over time, produce a sufficient number of blocky, prismatic, or columnar peds in B horizons stable enough for recognition of at least a weak grade of structure. These processes are related to shrink–swell phenomena associated with wetting and drying cycles. In soils with low shrink–swell potential (Southard and Buol 1988b), the strongest grades of blocky structure appear to occur in horizons that are saturated about 75% of the time and then are desiccated (Southard and Buol 1988a).

During the periods of desiccation, soil compression takes place because of the capillary tension of the water in the soil voids (Terzaghi and Peck 1967). The forces of capillary tension cause cracks to form in the soil mass, the walls of which later become ped faces. As the soil in this zone again becomes water saturated, it swells. The protruding points on the incipient ped faces make contact first, and shearing forces are focused at these points of contact. This causes coarser particles, such as sand grains, to move away from the pressure zone, leaving the platy clay particles that are forced into a parallel orientation on the ped face by the pressure. These modifications cause the soil to crack in the same place upon subsequent drying, thereby increasing the stability of the peds and resulting in recognizable blocky structure that persists (White 1966). These same shrink-swell processes can contribute to the conversion of rock fabric to soil fabric (Frazier and Graham 2000). Wetting and drying without the confining weight of overlying soil seem to favor prismatic or columnar structure (e.g., Figure 10.6B).

Granular structure is usually present in surface horizons where biological factors have the greatest impact. Granular structural units may consist of fecal casts of soil fauna, such as earthworms, or they may be composed of soil material bound by fungal hyphae and very fine roots (Oades 1993). Stability of surface soil aggregates is dynamic but is apparently related to the amount and form of organic residues (Layton et al. 1993). Platy structure is also typically found in surface horizons and is caused either by compaction from surface traffic, or by processes involved in the formation of the vesicular layer in desert soils. (See Chapter 10, Aridisols.) Platy structure is sometimes found in petrocalcic horizons and duripans (Plate 10.2 q1), where it may be caused by vertical compressional forces that accompany the precipitation of secondary calcite and silica (Chadwick and Graham 2008).

Haploidization Processes. Pedoturbation is the process of mixing in the soil. A certain amount of mixing takes place in all soils. Eight kinds of pedoturbation, referred to as turbation by several authors, are recognized.

Faunal pedoturbation is soil mixing by animals such as ants, earthworms, moles, and rodents (Plate 14.1), and by human activity such as in tillage operations. Faunal pedoturbation is instrumental in the development of mollic epipedons. (See Chapter 15, Mollisols.) *Floral pedoturbation* is mixing by plants as in tree tipping that forms pits and mounds. Mixing processes by plants and animals are more generally referred to as *bioturbation. Congellipedoturbation,* perhaps better known as *cryoturbation*, is mixing by freeze–thaw cycles as in patterned ground of tundra and alpine landscapes. (See Chapter 12, Gelisols.) *Argillipedoturbation* is mixing of materials in the solum by shrink and swell movements of expansible clays as they wet and dry in the water-exchange cycles within the soil. (See Chapter 19, Vertisols.) *Aeropedoturbation* is mixing by movement of gases in the soil during and after rains as air in the soil pores is trapped by the saturated front of percolating water and compressed. *Aquapedoturbation* is mixing by currents of water cascading into cracks open to the surface. *Crystal pedoturbation* is mixing by growth of crystals, such as halite (NaCl). *Seismopedoturbation* is mixing by vibrations, notably earthquake tremors.

Pedoturbation by biotic and physical agents can destroy argillans, embed fragments of them in the soil matrix, or even move clay back up into the A horizon. Microscopic fragments of argillans have been observed in thin sections of material from large ant mounds (Baxter and Hole 1967). More dramatically, mixing by argillipedoturbation can dominate all other processes and determine the major characteristics of some soils like the Vertisols. (See Chapter 19.) Cryoturbation has very significant impact on the properties of some Gelisols (Chapter 12) and probably plays a significant role in forming some features in all soils that are periodically frozen.

Simplified Model of Solum Development

Consideration of a simplified theoretical model can give us an intimation of the complexities of soil genesis (Van Wambeke 1972, 1976). The model presented here is of an open system with respect to water and some soluble products of weathering of feldspar and biotite, but it is a closed one in regard to other materials. Organic matter is not considered in the calculations. We assume that the soil has developed from a column of homogeneous material originally like the current C horizon.

Coarse sand grains of resistant minerals are used as "index minerals." Because of their large size, the coarse sand grains of the index minerals (hereafter referred to as IM) are assumed to have been unaffected during pedogenesis. They have moved only insofar as collapse or expansion of entire horizons has raised, lowered, and separated them. We can calculate gains and losses of mineral matter for the first three horizons by comparing the weight of gains per cubic centimeter in a horizon with their weight in the C horizon. Increase or decrease of populations of IM grains per unit volume in the solum indicates, respectively, collapse or expansion of the soil material as compared with the C horizon.

The horizon designations of our model and corresponding data are shown in Table 5.2. In this model, we have designated four horizons of equal thickness to simplify calculations. In practice, the bulk density of each horizon is determined; the values in Table 5.2 have been fabricated to simplify calculations so that all measurements can henceforth be made on a weight basis. By optical- and particle-size distribution methods, the index mineral, nonclay, and clay contents are determined for each horizon (columns 5, 7, 8, respectively). The IM factor (column 6) is computed by setting the IM content in each horizon (column 5) equal to the IM content of the C horizon, that is, the assumed composition for the entire profile prior to soil formation. By multiplying the C horizon values of present total weight, clay, and nonclay contents (columns 4, 7, and 8) by the IM factor (column 6), we obtain a value for the original contents in each horizon (columns 9, 10, and 11). To determine the nonclay content change in each horizon (column 12), columns 8 and 11 are compared. Note that in the example, 20 g of nonclay are lost from the profile, most of it from the upper horizons. A similar comparison of columns 7 and 10 calculates the change in clay content. In the example, the nonclay loss equals the clay gain. When, in actual calculations, the clay formed is less than nonclay loss, erosion or leaching losses can be calculated. Also, in the example, column 14 (sum of columns 12 and 13) estimates the net clay translocation within the profile, assuming that the clay is formed at the site of the nonclay loss. This type of IM model is easily constructed on spreadsheet software on personal computers, thereby removing the need for tedious calculations.

Table 5.2. A simplified model of solum development using an index mineral calculation

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)
Horizon	Thickness of 1-cm^2 Column (cm)	Bulk Density (g/cc)	Present Total, Oven-Dry Weight (g)	IM[a] Weight (g)	IM[a] Factor	Present Clay Oven-Dry Weight (g)	Present Nonclay Oven-Dry Weight (g)	Original Total, Oven-Dry Weight (g)	Original Clay Oven-Dry Weight (g)	Original Nonclay Oven-Dry Weight (g)	Change in Nonclay Content (g)	Change in Clay Content (g)	Net Clay Transocation (g)
A	20	1.0	20	3.2	1.6	5	1.5	40	8	32	−17	−3	−20
Bt1	20	1.5	30	1.2	0.6	20	10	15	3	12	−2	+17	+15
Bt2	20	1.25	25	1.6	0.8	10	15	20	4	16	−1	+5	+5
C	20	1.25	25	2.0	1.0	5	20	25	5	20	0	0	0
			100	8.0		40	60	100	20	80	−20	+20	0 loss

Note: Column 2 gives present-day thicknesses of the four soil horizons. Corresponding thicknesses of parent material at time zero were 32 cm, 12 cm, 16 cm, and 20 cm, respectively. Values in column 12 correspond to weights of new clay formed in each horizon during pedogenesis.

[a] IM = index mineral.

Real soils, unlike the theoretical model just demonstrated, are open to both gains and losses of materials and popular modeling approaches incorporate those aspects (Brimhall et al. 1991; Stolt et al. 1993). Although the model demonstrated here has been prepared for simplicity, it illustrates profile trends that are common in soils.

Perspective

Processes of soil formation constitute a complex suite or sequence of events. Many different processes may be taking place at the same time. Of the theoretically possible processes, only those for which the required reagents are present actually take place. Many reactions also require rather specific environmental conditions of temperature and water that may be lacking in some soils. What we see in soils is the balance or net effect of all of these processes, though one or a few may dominate in a particular soil. The cumulative effects of previously operating soil processes may cause the soil to reach a pedologic threshold, causing a whole new suite of processes and resulting soil properties to occur. These processes involve additions to soil, losses from the soil, physical movement of constituents from one soil to another in the landscape, translocation of materials within a soil, and transformation of substances within a soil by decomposition and synthesis.

The net effect of these processes is to create and maintain identifiable soil morphology and composition within the existing dynamics of conditions at points on the earth's surface. The vast array of possible pedogenic processes acting on landscapes is bound to bring about alterations of all kinds of soil features locally. Those features that a pedologist observes either have been formed by the existing environments or were formed in a previous environment and have persisted.

Modern Soil Classification Systems

6

Soil classification systems abound both historically and currently. Various systems range in scope from unstructured names for soils among indigenous people within small areas to complex systems that attempt to organize soil information throughout the world. In this chapter we present the higher categories of a few systems to illustrate various objectives and concepts people have regarding soil classification. Some comparisons with *Soil Taxonomy* are made, but a more complete discussion of *Soil Taxonomy* is presented in Chapter 7 and subsequent chapters.

Logic and Cognitive Science in Soil Classification

Through use of the principles of logic (such as set forth by Mill [1925] and described by Cline [1949]), we have a frame of reference, a terminology, and a means for analyzing and understanding the relationships among objects such as soils. Classification of objects, also called categorization, can help us abstract knowledge and principles from everyday experience in what some believe to be an unstructured world. Logic is useful in the classification process for studying relations among soils and placing them in classes according to prescribed rules. The categorization process—designing the categories—establishes differences and similarities among soils for most efficient information storage, retrieval, and communication. Cognate science, an understanding of genetic relationships among the objects classified, aids in structuring categories within a classification system.

Classification systems prepared by people differ for several reasons. However, the two general principles that guide the formation and arrangement of categories in classification systems follow:

1. Principle of cognitive economy. The role of classification is to provide maximum information with the least cognitive effort (knowledge and recall) while reducing infinite differences among objects being classified to useful and workable proportions.
2. The principle of perceived population structure. The world we perceive comes as structured information rather than with arbitrary attributes that must be sorted out by logic. When faced with a set of objects, people tend to, and can, sort them into clusters to reduce the information load and facilitate processing

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

information about them. The clusters they select will be according to natural breaks in similarity and dissimilarity. Categories are formed to be maximally differentiable from each other and have maximum "cue validity" (Rosch and Lloyd 1978; Rosch et al. 1976), that is, the categories or classes with the most attributes common to members of that category (class) and the least number of attributes shared with members of other categories (classes).

Other principles and theories of cognitive science relative to categorization and classification important and relevant for classifying soils follow:

3. The expertise effect in categorization. Classes and categories are classified according to differing sets of attributes. Each person will tend to use differing sets of criteria for defining classes and categories depending on their experience and training (Rosch et al. 1976).
4. Culture and environment effect how people form categories of similar objects. The correlation structures (properties that are assumed highly related and are assumed to be important diagnostic and differentiating criteria) are modified by "selective ignorance" and "exaggeration of attributes" by each classifier and are mirrored in their classification systems (Rosch et al. 1976). We shall see these principles displayed later in this chapter when we discuss soil classification systems of various countries and people.
5. The classifier's observations of samples or specimens (exemplars) lead to the development of prototype concepts. These prototype concepts then tend to represent the central tendency of a category or class and provide the basis for classification judgments.

The geographic distribution of soil precludes uniform comprehension of all soils by any one or group of soil scientists. Therefore, some classification systems tend to reflect concern and expertise for only a limited range of soil properties within the total population of soils in the world. The geographic confinement of soil scientists has been somewhat removed as more rapid travel and communication became available and newer classification systems tend to be more evenly inclusive of all soils. There are also differences in philosophies that guide the structure of soil classification systems. Many soil classification systems emphasize concepts and theories of soil genesis that identify soils as an independent entity related to other ecosystem components while other systems emphasize soil properties that quantitatively reflect how soils affect soil use (Buol 2003). These differences are reflected among the following abstracts of soil classification efforts by different groups of soil scientists.

Indigenous Soil Classifications

Indigenous, often termed 'folk' soil and land classification exists among localized communities throughout the world. These classifications essentially have no formal categorical structure and are limited in geographic scope to rather limited areas wherein undisturbed vegetation (ecosystems) is observed or human food is grown by similar cultivation practices. Many indigenous soil names have been preserved in more formal national and international soil classification systems. Tabor and Krasilnikov (2002) have published a worldwide list of more than 1,000 indigenous soil and landscape names, with brief descriptions and approximate correlation to the World Reference Base (FAO, ISRIC, and ISSS 1998) names.

We do not attempt to discuss indigenous classifications but strongly encourage students to familiarize themselves with the informal soil names and the soil characteristics associated with those names within any area they traverse. Local citizens residing in an area experience how their lives are affected by various soils within the geographic range of their experience and can often relate dynamic soil changes that occur seasonally or as a result of sporadic weather events. This is especially applicable to subsistence farmers in remote areas where the wellbeing of their families depends on the food they can produce on the various soils in the area. Therefore, they incorporate information that even the most accomplished soil scientist often cannot become aware of during a short-term study of the area.

By familiarization through conversations with local farmers and examination of the various soils they identify by indigenous names, a soil scientist is able to relate locally important soil and landscape characteristics to formalized soil names and/or map unit names. Also, if soil reports and/or maps are to be used by others seeking to communicate with the local people, the incorporation of indigenous names is very helpful. When indigenous people find familiar soil names, correctly used in a report or on a soil map, they have greater confidence in that work. As scientists are able to identify soil characteristics of local concern and relate those characteristics to characteristics identified by criteria of more formal classification systems, they are able to more successively transfer soil management information and technologies that have been successful in other areas with similar soils but unknown locally.

Regional and National Soil Classification Systems

We have selected a few examples of soil classification systems that historically have influenced more modern systems and a few national systems that illustrate ongoing soil classification endeavors. Students should not be overly concerned by the number and contrasting nature of the various systems, but keep in mind that the objectives of any classification system is to facilitate communication of knowledge with an intended audience (Buol 2006). Thus, national systems are most often designed to incorporate only soils that occur within that nation and convey only information considered pertinent to the concerns of citizens and scientists therein. Approaches and concepts of all soil classification systems take place as soil science develops new technology with which to characterize soil.

Early Soil Classification in Europe. Pedology in Europe has been strongly influenced by the Russian pioneers in pedology and by certain European soil scientists who emphasized or extensively used chemical methods of differentiating soils (Ramann 1911, 1918; Sigmond 1938). In addition to zonal concepts of climate and vegetation, laboratory-derived parameters such as silica/sesquioxide (Si/Fe_2O_3 + Al_2O_3) and silica/alumina ratios, determined from total elemental analyses were used as differentiating characteristics. Saturated conditions strongly affect agricultural uses of soils, and a high degree of significance was attached to wet (hydromorphic) soils in most European classifications. Most soils in Western Europe are formed in Late Pleistocene glacial deposits and have limited clay translocation. European soil scientists generally placed less emphasis on Bt horizon formation than soil scientists in the United States. Soils surrounding villages in Europe have been intensively cultivated, received almost continuous applications of manure since medieval times, and thus reflect the imprint of human activities more than soils in the United States. Many areas of Southern Europe are ecologically different from most of the United States in that they either have or approach a Mediterranean climate, with cool, moist winters and hot, dry summers.

The classification system of Kubiena (1953) was developed within the context of soils in Europe but was designed as a system of potential worldwide application. The system has a theme of soil developmental paths upon which conceptual classes are based. It also has a system of keys, based on Kubiena's premise that even though the nature of a soil is the total of all of its properties, it can be distinguished from associated soils by one critical differentiating characteristic (Butler 1980). Differentiating characteristics used include type of profile, chemical properties, humus type, pedogenic processes, and micromorphology.

The divisions, classes, and types of Kubiena's system are shown in Table 6.1.

Soil Classification in Russia (1997). Soil classification in Russia and the former Soviet Union historically influenced several classification efforts and continued the general approach first outlined by Dokuchaev and Sibirtsev. This included a strong genetic emphasis on evaluating soil properties to pedogenic processes and soil-forming factors. The action of soil-forming factors to produce soil properties in kinds of profiles called "soil types," or genetic soil types was strongly emphasized (Gerasimov and Ivanova 1959).

The first approximation of a new soil classification system was published in 1982 (Fridland 1982). After several stages of development, the new system was published in 1997. English translations of the 1997 Russian soil classification system provide an overview of the most recent classification trends in Russia (Gerasimova 2001; Karasilnikov 2002). A hierarchical structure provides eight categories: Trunk, Section, Type, Subtype, Genus, Class, Subclass, and Sort. The highest three categories are presented in Table 6.2.

The highest category, Trunk, recognizes three taxa: Postlithogenic (soils formed on previous parent material with negligible modern accumulation on the surface),

Table 6.1. Natural system of Kubiena (1953)

- A. Subaqueous or Underwater Soils
  - AA. Subaqueous soils not forming peat
    1. Protopedon (subaqueous raw soil)
    2. Dy
    3. Gyttja
    4. Sapropel
  - AB. Peat-forming subaqueous soils
    5. Fen
- B. Division of Semiterrestrial or Flooding and Groundwater Soils
  - BA. Semiterrestrial raw soils
    6. Rambla (raw warp soil)
    7. Rutmark (Arctic and snow basin)
    8. Raw gley soil
  - BB. Anmoorlike soils
    9. Anmoor
    10. Marsh
  - BC. Semiterrestrial peat soils
    11. Carr (Transition Wood Moor)
    12. High Moor
  - BD. Salt soils
    13. Solonchak
    14. Solonetz
    15. Solod
  - BE. Gley soils with land humus formation
    16. Gley
    17. Gray warp soils
    18. Rendzinalike warp soils
    19. Smonitza or Chernozemlike warp soils
    20. Vega
- C. Terrestrial or Land Soils
  - CA. Terrestrial raw soils
    21. Raw soils of cold deserts
    22. Dry desert raw soils
    23. Syrozem (raw soils of temperate zones)
  - CB. Rankerlike soils
    24. Ranker
  - CC. Rendzinalike soils
    25. Eurendzina
    26. Pararendzina
  - CD. Steppe soils
    27. Serozem
    28. Burozem (Brown Desert steppe soils)
    29. Kastanozem (Chestnut-colored soil)
    30. Chernozem
    31. Para-chernozem
    32. Para-serozem
  - CE. Terrae calxis
    33. Terra (includes Terra Fusca, Terra Rossa, etc.)

continues

Table 6.1. *Concluded.*

CF.	Boluslike silicate soils
	34. Braunlehm (Brown loams)
	35. Rotlehm (Red loams)
CG.	Latosols
	36. (Lateritic) Roterde
CH.	Brown earths
	37. Braunerde
CI.	Pseudogley
	38. Pseudogley
CJ.	Podzol
	39. Semipodzol
	40. Podzol

Synlithogenic (soils formed with periodical or continuous accumulation of new material on the surface), and Organogenic (soil formation on organic material). The Section category groups soils that have similar characteristics considered to be formed by combinations of soil-forming processes that include a Section of "cultivated soils, previously affected with intensive erosional processes," that is, Agrobrazems. The Type category identifies a number of color characteristics and of particular interest identifies somewhat disturbed by agricultural cultivation as "Argic" Types. The system continues to outline five lower categories that place strong emphasis on human modification of soil properties. Climatic conditions and mineralogical composition are classified independently.

Soil Classification in the Netherlands. The Netherlands is a small, intensely agrarian country with a history of soil classification dating back to the mid nineteenth century. Between 1964 and 1995, 1:50,000 scale soil maps of the entire country were made and the current soil classification system forms the legend for those maps. About half of the country is below sea level, protected by dikes and dunes. Most soils are formed from marine or fluvial sediments, Pleistocene glacial deposits, loess, or are organic. Many soils require engineered drainage to facilitate cultivation. About 40% of the soils are sandy and 25% are organic (peat). No soils are formed from consolidated rock (Hartermink and de Bakker 2006).

The current soil classification system consists of four categories: order, suborder, group, and subgroup, however additional categories are used to identify map units on 1:50,000 scale soil maps. The order names define Peat, >40 cm organic soil material; Podzol, presence of a spodic horizon; Brick, presence an argillic horizon; and Vague, lacking characteristics of the other orders. Hydromorphic properties ('Hydro') are differentiated in the suborder name, and textures are most often expressed in the group name. Interestingly, although not presented in Table 6.3, subgroup names are most often formed from the names of villages, rivers, or the dominant type of land use. Considering the small size of the country, these names, although having distinct and defined soil properties, can be considered indigenous and provide for good communication with citizens.

Table 6.2. Russian Soil Classification

Trunk	Section	Type
Postlithogenic	Gleyzems	Gleysems
		Peat gleyzems
		Argic gleyzems
		Argic peat gleyzems
	Cryozems	Cryozems
		Peat cryozems
	Al-Fe humus soils	Podburs
		Dry podburs
		Gleyic peat podburs
		Sod Al-Fe-humus
		Sod Al-Fe-humus gleyic
		Podzols
		Dry peat podzols
		Sod podzols
		Gleyic podzols
		Gleyic peat podzols
		Gleyic sod podzols
		Argic sod podzols
		Argic peat gleyic podzols
	Texture differentiated soils	Podzolic
		Gleyic podzolic
		Peat gleyic podzolic
		Sod podzolic
		Sod gleyic podzolic
		Grey
		Gleyic gray
		Dark-humus podbels
		Dark humus gleyic podbels
		Argic sod podzolic
		Argic sod gleyic podzolic
		Argic peat gleyic podzolic
		Argic grey
		Argic gleyic grey
		Argic dark-humus podbels
		Argic dark-humus gleyic podbels
	Svetlozems	Svetlozems
		Iron-illuviated svetlozems
		Sod svetlozems
	Organic matter accumulating soils	Humic (sod)
		Dark-humus
		Rendzinas
		Mud
		Mud rendzinas
		Argic humic
		Argic mud
		Argic rendzinas
	Metamorphic soils	Burozems (brown soils)
		Raw humus burozems

continues

Table 6.2. *Continued.*

Trunk	Section	Type
		Pale
		Argic pale
	Humus-accumulating soils	Chernozems
		Clay-illuvial chernozems
		Black compact
		Chernozem-like
		Chestnut
		Argic chernozems
		Argic clay-illuvial chernozems
		Argic compact
		Argic chernozem-like
		Argic chestnut
	Humus accumulating hydrogenically transformed soils	Humus gley
		Mud gley
		Humus crypto-gley
		Mud crypto-gley
		Argic mud gley
		Argic humus crypto-gley
	Low-humus carbonate-accumulating soils	Arid brown
	Alkaline clay-differentiated soils	Dark solonetz
		Light solonetz
		Dark crypto-gleyic solonetz
		Light crypto-gleyic solonetz
		Argic dark solonetz
		Argic light solonetz
		Argic dark crypto-gleyic solonetz
		Argic light crypto-gleyic solonetz
		Light (sod) solod
		Dark solod
		Light sod gleyic solod
		Dark crypto-gleyic solod
		Peat gleyic solod
		Argic light sod solod
		Argic dark solod
		Argic sod gleyic solod
		Argic dark crypto-gleyic solod
		Peat gleyic solod
		Argic dark solod
		Argic sod gleyic solod
		Argic dark crypto-gleyic solod
	Holomorphic soils	Light solonchaks
		Dark solonchaks
		Light gleyic solonchaks
		Dark crypto-gleyic solonchaks
		“sorovey” solonchaks
		Secondary solonchaks
	Lithozems	Peat lithozems
		Raw humus lithozems
		Light lithozems

Table 6.2. *Continued.*

Trunk	Section	Type
		Dark lithozems
		Argic light lithozems
		Argic dark lithozems
	Abrazems	Al-Fe humus abrazems
		Clay-illuvial abrazems
		Metamorphic abrazems
		Carbonate-accumulative abrazems
		Solonetzic abrazems
	Agrozems	Light agrozems
		Dark agrozems
		Light gleyic agrozems
		Dark crypto-gleyic agrozems
		Peat agrozems
		Peat-mineral agrozems
		Al-Fe humus agrozems
		Texture-differentiated agrozems
		Metamorphic agrozems
		Metamorphic agrozems
		Clay-illuvial agrozems
		Carbonate-accumulative agrozems
		Light solonetz agrozems
		Dark solonetz agrozems
	Agroabrazems (agroerozems)	Agroabrazems
		Gleyic agroabrazems
		Crypto-gleyic agroabrazems
		Al-Fe humus agroabrazems
		Metamorphic agroabrazems
		Clay-illuvial agroabrazems
		Carbonate-accumulative agroabrazems
		Solonetzic agroabrazems
Sinlithogenic soils	Weakly developed soils	Stratified alluvial soils
		Stratified ash volcanic
		Aeolian
		Proluvial
	Alluvial soils	Light–humus (sod) alluvial
		Dark-humus alluvial
		Alluvial peat-gleyic
		Alluvial mud-gleyic
		Alluvial sod-gleyic
		Alluvial humus-gleyic
		Alluvial humus-crypto-gleyic marl
		Alluvial humus-gleyic iron-ore
		Alluvial compact
		Argic alluvial light-humus (sod)
		Argic alluvial dark-humus
		Argic alluvial peat-gleyic
		Argic alluvial mud-gleyic
		Argic alluvial sod-gleyic

Table 6.2. *Concluded.*

Trunk	Section	Type
		Argic alluvial humus-gleyic
		Argic alluvial humus-crypto-gleyic
		Argic alluvial humus-crypto-gleyic marl
		Argic alluvial humus-gleyic iron-ore
		Argic alluvial compact
	Alluvial agrozems	Light alluvial agrozems
		Dark alluvial agrozems
		Gleyic alluvial agrozems
		Crypto-gleyic alluvial agrozems
		Peat-mineral alluvial agrozems
	Volcanic soils	Ochrous
		Dry peat ochrous
		Ochrous podzolic
		Dry peat ochrous podzolic
		Argic ochrous
	Stratozems	Light stratozems
		Dark stratozems
		Light stratozems on a fossil
		Dark stratozems on a fossil
		Argic light stratozems
		Argic dark stratozems
		Argic light stratozems on a fossil
		Argic dark stratozems on a fossil
Organogenic soils	Peat soils	Peat oligotrophic
		Peat eutrophic
		Dry peat
	Argic peat soils	Argic oligotrophic peat
		Argic eurtrophic peat
		Argic

Source: Krasilnikov 2002.

Pedology in New Zealand. New Zealand soil scientists made significant contributions to the development of the order Andisols in *Soil Taxonomy*. However, unsatisfied with the slow response of the United States to recognize alterations, particularly to the classification of important Inceptisols in New Zealand, a separate national system to include only the soils in New Zealand was developed (Hewitt 1992, 1998). One principle of the new national system is to maintain correlation with *Soil Taxonomy* while providing more rapid incorporation of detailed field data. Nomenclature to bridge the gap between the high-precision and expensive measurements required by *Soil Taxonomy* and the less laboratory-determined, but reproducible, field measurements usable for land-system modeling was desired.

The current system identifies several soils that are extensive within New Zealand at the order level that are only recognized in lower categories in *Soil Taxonomy*. The system

Table 6.3. Three highest categories of the Netherlands soil classification

Order	Suborder	Group
Peat soils	Earthy peat soils	Clayey earthy peat
		Clay-poor earthy peat
	Raw peat soils	Initial raw peat
		Ordinary raw peat
Podzol soils	Moder podzol soils	Moder podzol
	Hydropodzol soils	Peaty podzol
		Ordinary hydropodzol
	Xeropodzol soils	Xeropodzol
Brick soils	Hydrobrick soils	Hydrobrick
	Xerobrick soils	Xerobrick
Earth soils	Thick earth soils	"Enk" earth
		"Tuin" earth
		Peaty earth
	Hydroearth soils	Sandy hydroearth
		Clayey hydroearth
	Xeroearth soils	Sandy xeroearth
		Clayey xeroearth
Vague soils	Initial vague soils	Initial vague
	Hydrovague soils	Sandy hydrovague
		Clayey hydrovague
	Xerovague soils	"Krijt" xerovague
		Sandy xerovague
		Clayey xerovague

Source: Hartemink and de Bakker 2006. Translation of H. deBaker and J. Schelling. 1989. *Systeem van Bodemelassificatie voor Nederland, de Hogre Niveaus;* PUDOC Wageningen, 1989 (in Dutch, with English summary).

defines 15 orders (Table 6.4). Soils formed from parent materials of volcanic origin, so extensive in New Zealand, are identified in two orders, Allophanic and Pumice soils. No quantitative criteria identify soil moisture and temperature regimes, and soils altered to a depth of 30 cm or more by human activity are placed in a separate soil order.

Soil Classification in Canada. The Canadian soil classification system is designed to classify only soils known to exist in Canada. It is a hierarchical taxonomy with five categories: order, great group, subgroup, family, and series. In 1998 (Soil Classification Working Group 1998), one new order (Vertisolic) and four great groups were added to the previous system (Canada Soil Survey Committee, Subcommittee on Soil Classification 1978). The taxa are differentiated by observable and measurable soil properties with sufficient genetic structure in the higher categories to provide a theoretical basis for distinguishing among the 10 orders and 31 great groups. Only about 10 percent of the soils in Canada are likely to be cultivated; thus, diagnostic criteria include all horizons to the surface. Classes and subclasses of particle size,

Table 6.4. Abbreviated key to soil orders in New Zealand

Key Criteria (Follow in sequence)	Soil Order
Soils with 30 cm or more of organic soil material within 60 cm of the surface	Organic Soils
Other soils with reductimorphic grayish colors and no fragipan or duripan	Gley Soils
Other soils with pH values less than 5.5 between base of A horizon and 60 cm	Ultic Soils
Other soils with either a podzolic B horizon or an ortstein-pan	Podzols
Other soils with 35 cm or more allophonic soil material within 60 cm of surface	Allophanic Soils
Other soils with 35 cm or more vitric soil material over a weathered B horizon	Pumice Soils
Other soils with A horizon of 3 or less moist color value and sometimes cracks to a depth of 30 cm	Melanic Soils
Other soils with a weathered or calcareous B horizon and no fragipan	Semiarid Soils
Other soils with the upper boundary of a 30 cm or thicker oxidic horizon within 20 and 60 cm of the surface	Oxidic Soils
Other soils with a clayey textured B horizon that is moderately of strongly pedal	Granular Soils
Other loamy fine sand or finer soils with (specific) light colors below the A horizon	Pallic Soils
Other soils with a weathered B horizon 10 or more cm and (specific) colors	Brown Soils
Other soils with 30 cm or more deposited or mixed by people	Anthropic Soils
Other soils with a distinct topsoil, or weathered B horizon; less than moderate fluidity within 30 cm of surface	Recent Soils
Other soils	Raw Soils

Abstracted from Hewitt 1998.

mineralogy, depth, reaction, calcareousness, and soil temperature and soil moisture regimes are criteria at the family category. Table 6.5 presents an abbreviated key to the orders and a listing of great groups in each order. The complete system is available on-line at http://www.pedosphere.com/resources/cssc3rd.

Differentiating characteristics for each of the categories are as follows:

Order: soil properties reflecting the effects of the dominant soil-forming processes and the soil environment.

Great Group: properties reflecting the strength of dominant pedogenic processes in addition to the major one.

Subgroup: taxa based on congruence with the central concept of a group and the associated kind and arrangement of horizons.

Family: taxa separated on the basis of mineralogy, texture, and climatic factors.

Series: differentiated on the basis of morphology and arrangement of horizons within each family.

The Canadian system is similar to *Soil Taxonomy*. Both are multicategorical with ascending lines of generalization from the lowest to the highest level. However, the Canadian system was designed only to accommodate soils in Canada.

Soil Classification in Australia. The Australian continent has a wide range of climates, ranging from humid tropical in the north to extremely dry deserts in the

Table 6.5. Abbreviated keys in the Canadian soil classification system (1998)

Key Description	Order Name	Great Groups
Soils with permafrost within 1 meter	Cryosolic	Organic Cryosol Turbic Cryosol Static Cryosol
Other soils with more than 17% organic carbon to 40 cm or more	Organic	Folisol Fibrisol Mesisol Humisol
Other soils with both a vertic and slickenside horizon within 1 meter	Vertisolic	Vertisol Humic Vertisol
Other soils with a podzolic B and no Bt horizon within 50 cm	Podzolic	Humic Podzol Ferro-Humic Podzol Humo-Ferric Podzol
Other soils with saturated and reducing conditions sometime during the year (specific low chroma colors within 50 cm)	Gleysolic	Luvic Gleysol Humic Gleysol Gleysol
Other soils with a solonetzic B horizon	Solonetzic	Vertic Solonetz Solod Solodized Solonetz Solonetz
Other soils with a chernozemic A horizon	Chernozemic	Brown Chernozem Dark Brown Chernozem Black Chernozem Dark Grey Chernozem
Other soils with a Bt horizon	Luvisolic	Gray Brown Luvisiol Gray Luvisol
Other soils with a Bm, Btj, or Bfj horizon 5 or more cm thick	Brunisolic	Melanic Brunisol Eutric Brunisol Sombric Brunisol Dystric Brunisol
Other soils	Regosolic	Humic Regosol Regosol

Source: Soil Classification Working Group 1998.

interior and warm subhumid temperate climates in Southern and Western Australia. About one-third of Australia is arid and another third semiarid (Leeper 1950). Ground surfaces (geomorphic surfaces or landscapes) have a wide range in age. Therefore, it should not be surprising to find that climatic gradients, vegetation differences, and paleosols were given some emphasis in the many Australian soil classification systems.

Among the first attempts at soil classification in Australia was that of Jensen (1914), who presented a system based on geology of the initial soil material. Prescott (1931) prepared a general soil map of Australia and a classification system based on soil properties. He strongly emphasized the role of climate and vegetation, and introduced the work and ideas of Dokuchaev and established 18 generalized soil environment groups, with a central concept soil profile for each. With additional studies, it became apparent that not all soils in an environment group even approximately conformed to the generalized soil profile for that group hence the environmental grouping concept was abandoned.

Stephens revised Prescott's approach with more emphasis on soil features and developed a morphogenetic soil classification. This culminated in a classification with 47 great soil groups defined qualitatively, primarily on the basis of morphology of central concept profiles (Stephens 1962).

Northcote (1960) proposed a Natural Objective Classification, based on a bifurcating system (two classes per category), with specific values and limits for the properties of soils in each class. Soil properties were chosen to cover the degree and broad kind of soil profile differentiation, and genetic connotations were deemphasized.

Another system of soil classification was published by Stace et al. (1968). It has 43 great soil groups based on a general kind of profile and a higher category of classes differentiated according to broad generalized characteristics, such as "no profile differentiated" and "dark soils." This was a rather skeletal classification composed of two levels of "conceptual classes."

Northcote (1971) published an expanded key largely based on soil properties measurable in the field. The first (highest) separation was made on the basis of the soil texture profile. The second level of separation varies among the divisions established by the first split but includes color, consistence, and similar morphological properties. He made efforts to relate land use and landscape characteristics to the classes produced. Butler (1980) suggested that this system was rather incomplete and lacked a conceptual framework.

FitzPatrick (1971) published a rather different "bottom up" approach that started with characteristics of individual soil profiles and generalized with minimum genetic assumptions. The categories of the system are as follows:

Horizon pedounit: a combination of similar horizons (77 types of horizons are considered).

Group: a collection of pedounits that are similar to soil mapping characteristics equivalent to the U.S. soil series.

Subclass: groups placed together because of similarity in two or three kinds of horizons.

Class: the highest level of generalization.

Butler (1980) was critical of each of the previous systems and pointed out that although they provided a level of generalization for all soils, there were problems in dealing with soils at the lower levels.

Isbell (1992, 1993) cited difficulties with previous systems and developed a hierarchical system that envisioned only soils in Australia. The general philosophy was to create mutually exclusive classes at all levels by selecting differentia that reflected the most important variables within each class. More laboratory data were utilized as differentiae than in earlier Australian systems. The system outlines five

Table 6.6. Approximate key to orders in the Australian soil classification system

Key Description	Order Name
A. Soils resulting from human activities	ANTHROPOSOLS
B. Other soils with more than 40 cm of organic materials in upper 80 cm	ORGANOSOLS
C. Other soils with a Bs, Bhs or Bh horizon	PODOSOLS
D. Other soils with 35% or more clay in all horizons, cracks at least 5 mm wide opening sometime in most years and lenticular peds and/or slickensides in the solum	VERTOSOLS
E. Other soils with most of the profile saturated 2-3 months in most years	HYDROSOLS
F. Other soils with clear or abrupt textural B horizon that is strongly acid (pH_w, < 5.5)	KUROSOLS
G. Other soils with clear or abrupt textural B horizon with exchangeable sodium percentage of 6 or more	SODOSOLS
H. Other soils with a clear or abrupt textured B horizon with a pH of 5.5 or more	CHROMOSOLS
I. Other soils that are calcareous either directly below the Ap or Al horizon or throughout the solum	CALCAROSOLS
J. Other soils which have a moderate or stronger grade of structure and more than 5% Fe in the B2 horizon	FERROSOLS
K. Other soils which have a moderate or stronger grade of structure in the B2 horizon	DERMOSOLS
L. Other soils which have a massive or weak grade of structure in a B2 horizon with more than 15% clay	KANDOSOLS
M. Other soils with negligible pedological organization other than a minimal Al horizon	RUDOSOLS
N. Other soils	TENOSOLS

Source: Isbell 2002.

categories, with letter codes to identify each category, and employs several new morphological terms in addition to those used in the Australian Field Handbook (McDonald et al. 1990). Table 6.6 is an abbreviated key to the orders in that system (Isbell 1998, 2002).

Soil Classification in Brazil. Soils in the 850 million ha of Brazilian territory have undergone intensive investigation in the last 60 years, and their classification is being developed mainly by analyzing data from many reconnaissance and exploratory soil surveys. Older provisional systems have been published by Beinroth (1975) and Camargo et al. (1987). A new official national classification system was first released (EMBRAPA 1999) that currently is on its second approximation (EMBRAPA 2006). A generalized 1:5,000,000 scale soil map of Brazil based on the old system was presented by Camargo et al. (1981), and another map, based on the 1999 system, was published by IBGE/EMBRAPA (2001).

Brazilian soil scientists now recognize 13 classes of soils at the highest categorical level. Four categories were already outlined: order, suborder, great group, subgroup; families and series have yet to be defined. Some orders include soils that were previously separated, for instance the "Neossolos" that embraces soils formerly classed as "Regossolos," "Litossolos" and "Solos Litólicos," "Solos Aluviais," and "Areias Quartzosas" (Costa and Nanni 2006).

Many aspects of *Soil Taxonomy* were incorporated. The orders are differentiated mainly by presence or absence of well-defined diagnostic horizons, such as the "horizonte B latossólico" (latosolic B horizon, similar to oxic horizon) and "horizonte B textural" (similar to the argillic and kandic horizons). Several other diagnostic characteristics are considered for the suborders and great groups such as base saturation, clay activity, cation exchange capacity, effective cation capacity, sodium saturation, and some distinctive morphological characteristics. Subgroups are differentiated according to the central concept taxa for each great group and properties indicating characteristics of other categories or soils with extraordinary characteristics. However, soil moisture and soil temperature regimes are not considered.

Criteria to define the latosolic B horizon includes most criteria also used for the oxic horizon of *Soil Taxonomy*, but it also includes Ki and Kr weathering indexes (SiO_2/Al_2O_3 and $SiO_2/[Al_2O_3+Fe_2O_3]$ molar ratios, respectively, obtained from total analysis of the clay fraction). This diagnostic horizon is considered to be formed by pedogenic processes of intense loss of basis, silica, and residual concentration of aluminum and iron oxides; thus, besides the Ki and Kr indexes, color, base saturation, effective CEC, and iron oxide content are further used as suborder and great group differentiae in the "Latossolo" order.

Taxa names are derived primarily from the 1938 system of the United States (Baldwin et al. 1938), Kellogg's (1949) proposals for tropical soils, and the FAO/WRB system (FAO 1998). Soils in the Latossolos and Argissolos orders comprise about 75% of the country's area (IBGE/EMBRAPA 2001). The Latossolos are virtually the same as well-drained Oxisols of *Soil Taxonomy* without a kandic horizon, and the Argissolos are mainly Ultisols and low-activity clay Alfisols. In most older Brazilian reconnaissance soil maps, the class names were further quantified by phase names of "Distrófico" for base saturation less than 50% (as determined with soil CEC at pH 7), "Eutrófico" for base saturation equal to or above 50%, "Álico" if KCl extractable aluminum exceeds 50% of the effective cation exchange capacity (i.e., sum of exchangeable Al, Ca, Mg, K, and Na) or "Ácrico" if the effective cation capacity is very low (that is, less than $1.5\,cmol_c Kg^{-1}$ of clay). With so much of Brazil having soils with relatively low fertility and few morphological features that indicate weathering degree and soil fertility, Brazilian soil scientists have been more aggressive in incorporating chemical differentiate in their soil classification systems than most other countries.

In the current taxonomic system (EMBRAPA 2006), color is used as the main suborder criteria for the Latossolos and Argissolos (Table 6.7). For the Latossolos great groups, criteria such as base saturation, effective CEC, Fe oxide contents, and soil cohesion are used. The Álico criterion is now intended to be used as a family criterion.

Chinese Soil Classification. In 2001, under the project leadership of Professor Gong Zitong, the Cooperative Research Group of Chinese Soil Taxonomy published Chinese Soil Taxonomy (CRG-CST 2001), which updated the 1994 first proposal by Zitong (1994). The Chinese Soil Taxonomy is a six category hierarchical system based on quantified diagnostic horizons and diagnostic properties. Many of the 33 diagnostic horizons and 25 diagnostic properties, including soil temperature and soil

Table 6.7. Order and suborders in Brazilian soil classification

Ordem (Order) *Approximate Soil Taxonomy Equivalent (S)*	Subordem (Suborder)	Abbreviated Brazilian Criteria*
ARGISSOLOS *Ultisols and low clay activity Alfisols*	Bruno-Acinzentado	Darkening of upper B horizon
	Acinzentado	Grayish B horizon
	Amarelo	Yellowish B horizon
	Vermelho	Reddish B horizon
	Vermelho-Amarelo	Yellowish-red B horizon
	Vermelho-Amarelo	Yellowish-red B horizon
CAMBISSOLOS *Inceptisols*	Húmico	High carbon-content segment of Umbric epipedon (Humic)
	Flúvico	Erratic carbon distribution with depth (Fluvic properties)
	Háplico	Other
CHERNOSSOLOS *Mollisols*	Rândzico	Secondary calcium carbonate accumulation (Calcic)
	Ebânico	Dark colored subsurface horizon
	Argilúvico	Clay illuviation or eluvial horizon
	Háplico	Other
ESPODOSSOLOS *Spodosols*	Humilúvico	Illuviation of organic compounds with aluminum
	Ferrilúvico	Illuviation of iron compounds
	Ferrihumilúvico	Illuviation of organic compounds with aluminum and iron
GLEISSOLOS *Most Aquic suborders*	Tiomórfico	Sulfuric horizon or sulfidic materials
	Sálico	EC >7 dS/m within 100 cm
	Melânico	Thin (<40 cm) histic or umbric epipedon
	Háplico	Other
LATOSSOLOS *Non aquic Oxisols without kandic horizon*	Bruno	Brownish-colored B horizon (high elevation subtropics)
	Amarelo	Yellowish B horizon
	Vermelho	Reddish B horizon
	Vermelho-Amarelo	Yellowish-red B horizon
LUVISSOLOS *High clay activity Alfisols*	Crâmico	Bright B horizon (high-value, high-chroma color)
	Háplico	Other (dimmed B horizon)
NEOSSOLOS *Entisols*	Litólico	Lithic contact within 50 cm
	Flúvico	Erratic carbon distribution with depth (Fluvic properties)
	Regolítico	Lithic contact below 50 cm and high weatherable minerals
	Quartzarênico	Sandy particle-size class composed of resistant minerals
NITOSSOLOS *Oxisols with kandic horizon*	Bruno	Brownish-colored B horizon (high elevation subtropics)
	Vermelho	Reddish B horizon
	Háplico	Other
ORGANOSSOLOS *Histosols*	Tiomórfico	Sulfuric horizon or sulfidic materials
	Fólico	Folist (non water-saturated histic horizon)
	Háplico	Other

continues

Table 6.7. *Concluded.*

Ordem (Order) *Approximate Soil Taxonomy Equivalent (S)*	Subordem (Suborder)	Abbreviated Brazilian Criteria*
PLANOSSOLOS	Nátrico	Natric B horizon
Albic Ultisols and Alfisols (Albaqults Albaqualfs)	Háplico	Other
PLINTOSSOLOS	Pétrico	Petroferric contact or skeletal (>50%) iron-concretions layer
Plinthic and petroferric Oxisols, Ultisols, and Inceptisols	Argilúvico	Clay illuviation with plinthite concentration
	Háplico	Other
VERTISSOLOS	Hidromórfico	Redoximorphic properties
Vertisols	Ebânico	Dark-colored subsurface horizon
	Háplico	Other

Source: IBGE 2007, EMBRAPA 2006.
*Correlation with Soil Taxonomy and Brazilian Criteria Translated by Drs. Ricardo M. Coelho and Igo F. Lepsch.

Table 6.8. Abbreviated key to orders in Chinese soil taxonomy

Key Description	Order
Organic soil material	Histosols
Other soils with human made epipedons	Anthrosols
Other soils with a spodic horizon within 100 cm	Spodosols
Other soils with 60% or more andic soil properties	Andosols
Other soils with a ferralic horizon within 150 cm	Ferralosols
Other soils with vertic features within 100 cm	Vertosols
Other soils with aridic epipedon and certain other diagnostic horizons within 100 cm	Aridosols
Other soils with a salic horizon within 30 cm or an alkalic horizon within 75 cm	Halosols
Other soils with gleyic features within 50 cm	Gleyosols
Other soils with a mollic epipedons and 50% or more base saturation to 180 cm	Isohumosols
Other soils low activity clay-ferric horizon within 125 cm	Ferrosols
Other soils with an argic horizon within 100 cm	Agrosols
Other soils with a cambic horizon or certain other diagnostic horizons within 100 cm	Cambosols
Other soils	Primosols

Source: CRG-CST 2001.

moisture regimes, correspond to definitions in *Soil Taxonomy*. Additional diagnostic horizons characteristic of soils that have, for perhaps thousands of years, been used for flooded (paddy) rice cultivation are named and defined.

Only keys to the four highest categories, order, suborder, group, and subgroup are in the cited publication, but provisions are made for a family category, using particle size classes, mineral composition, and soil temperature regimes, and a series category using ‘old names’ independent of higher category nomenclature.

There are 14 orders (Table 6.8) and 39 suborders (Table 6.9) in the Chinese system. To reduce confusion, order names are spelled the same if their definition corresponds

Table 6.9. Suborders in Chinese soil taxonomy

Order Name	Suborders	Abbreviated Suborder Criteria
Histosols	Permagelic	Permafrost or permagelic temperatures within 200 cm
	Orthic	Other Histosols
Anthrosols	Stagnic	Anthrostagnic moisture regime
	Orthic	Other Anthrosols
Spodosols	Humic	60 g kg^{-1} or more organic carbon in 10 cm of spodic layer
	Orthic	Other Spodosols
Andosols	Cryic	Cryic or colder soil temperature regime
	Vitric	Coarser than silt loam in all horizons within 100 cm
	Udic	Other Andosols
Ferralosols	Udic	Have a udic soil moisture regime
Vertosols	Aquic	Have an aquic soil moisture regime
	Ustic	Have an ustic soil moisture regime
	Udic	Have a udic soil moisture regime
Aridosols	Cryic	Have a cryic soil temperature regime
	Orthic	Other Aridosols
Halosols	Alkalic	Have an alkalic horizon within 75 cm
	Orthic	Other Halosols
Gleyosols	Permagelic	Have permafrost within 200 cm
	Stagnic	Have Stagnic soil moisture regime
	Orthic	Other Glyosols
Isohumosols	Lithomorphic	Have lithologic character of coral sand or carbonate rocks
	Ustic	Have an ustic soil moisture regime
	Udic	Other Isohumosols
Ferrosols	Ustic	Have an ustic soil moisture regime
	Perudic	Have a perudic soil moisture regime
	Udic	Have a udic soil moisture regime
Agrosols	Boric	Have frigid or cryic soil temperature regime
	Ustic	Have an ustic soil moisture regime
	Perudic	Have a perudic soil moisture regime
	Udic	Have a udic soil moisture regime
Cambosols	Gelic	Have cryic or colder soil temperature regime
	Aquic	Have an aquic soil moisture regime
	Ustic	Have an ustic soil moisture regime
	Perudic	Have a perudic soil moisture regime
	Udic	Other Cambosols
Primosols	Anthric	Have anthroturbic layer within 50 cm
	Sandic	Have sandy deposits with lithologic characteristics
	Alluvic	Have other deposits with lithologic characteristics
	Orthic	Other Primosols

Source: CRG-CST 2001.

to internationally accepted definitions, and new names are constructed for orders in which definitions differ from other taxonomies. Approximate equivalents comparing the Chinese orders to orders in *Soil Taxonomy* and groups in the World Reference Base are presented in tables. The four highest categories utilize the principle of segmental-continuous names greatly facilitating association of subgroups to all higher categories. For example the subgroup name "Lithic Udic-Vitric Andosols" contains the group name "Udic-Vitric Andosols," the subgroup name "Vitric Andosols," the suborder name and the order name "Andosols."

World Reference Base (WRB)

This international effort to provide a reference base that could incorporate the numerous soil classification systems into a system of names that would be universally recognized was started in 1980 and carried forward by working groups. The effort is an outgrowth of a two-categorical system (Dudal 1968a, 1968b) to define the map units of the FAO/UNESCO world soil map project (FAO 1988).

During and following the construction of the world soil map, the need for more detailed criteria became evident, and several diagnostic soil horizons, materials, and properties were defined, and additional soil groups and subunits identified for international use were published (FAO, ISRIC, and ISSS 1998). More recently the working group of the International Union of Soil Sciences (IUSS) published an update of that system (IUSS Working Group WRB 2006). Among the basic principles adopted by the working groups is an attempt to define and use diagnostic soil characteristics related to soil-forming processes and diagnostic features of significance to soil management as criteria for classifying soils (Table 6.10). As a basic principle, they have decided not to incorporate climatic parameters (soil temperature and soil moisture regimes of *Soil Taxonomy*) as classification criteria.

The WRB classification system consists of two tiers. The top tier or category consists of 32 groups designed for naming map units on small-scale maps such as the FAO/UNESCO world soil map. Groups are defined by specific diagnostic horizons and/or properties. An abbreviated key to the reference top tier soil groups is presented in Table 6.11.

For second tier classification, each group has a list of prefix and suffix qualifiers that may be used to more specifically identify the group. Prefix qualifiers, placed before the group name, typically identifies a property or properties that intergrade toward another group. Suffix qualifiers, placed in brackets following the group name, identify specific soil chemical, physical, or textural properties not commonly associated with the characteristics of the group. Second tier classification is designed for use in naming soil map units on more detailed, larger scale soil maps.

The diagnostic horizons, properties, and materials of the World Reference Base are reasonably well coordinated with diagnostic horizons and materials in *Soil Taxonomy* (Table 6.10). However, correlation or harmonization of WRB groups with orders in *Soil Taxonomy* is problematic because soil moisture regimes are not used in the WRB classification (Buol et al. 2006).

Table 6.10. Diagnostic horizons, properties, and soil materials of World Reference Base and approximate *Soil Taxonomy* equivalent or description

Diagnostic Horizons	Approximate *Soil Taxonomy* Equivalent or (Description)
Albic horizon	Albic materials
Anthraquic horizon	(Puddled layer and plow pan)
Anthric horizon	(Ap horizon)
Argic horizon	Argillic horizon
Calcic horizon	Calcic horizon
Cambic horizon	Cambic horizon
Cryic horizon	Permafrost
Duric horizon	(10% or more silica cemented Durinodes)
Ferralic horizon	Oxic and Kandic horizons
Ferric horizon	(Coarse red mottles)
Folic horizon	Folistic epipedon
Fragic horizon	(Strong structure, restricts roots and water movement to cracks)
Fulvic horizon	Andic soil properties
Gypsic horizon	Gypsic horizon
Histic horizon	Histic epipedon
Hortic horizon	Anthropic epipedon
Hydragric horizon	(Redox features resulting from wet cultivation)
Irragric horizon	(Mineral surface horizon resulting from irrigation)
Melanic horizon	Melanic epipedon
Mollic horizon	Mollic epipedon
Natric horizon	Natric horizon
Nitric horizon	(Argillic or Kandic horizon with >30% clay and shiny ped faces)
Petrocalcic horizon	Petrocalcic horizon
Petroduric horizon	Duripan
Petrogypsic horizon	Petrogypsic horizon
Petroplinthic horizon	Petroferric contact
Plaggic horizon	Plaggen epipedon
Plinthic horizon	Plinthite
Salic horizon	Salic horizon
Sombric horizon	Sombric horizon
Spodic horizon	Spodic horizon
Takyric horizon	(Clayey surface crust on arid soils periodically flooded)
Terric horizon	(Mineral material applied by humans)
Thionic horizon	Sulfuric horizon
Umbric horizon	Umbric epipedon
Vertic horizon	(30% or more clay and slickensides)
Voronic horizon	(Black, 80% or more B.S. [CEC_7] earthworm-rich epipedons)
Yermic horizon	(Surface layer of gravel, desert pavement)
Diagnostic Properties	
Abrupt textural change	Abrupt textural change
Albeluvic tonguing	Interfingering of Albic materials
Andic properties	Andic soil properties
Aridic properties	(Surface features resulting from wind)
Continuous rock	Lithic contact
Ferralic properties	(Apparent CEC_7 <24 cmol kg^{-1} clay)

continues

Table 6.10. *Concluded.*

Diagnostic Horizons	Approximate *Soil Taxonomy* Equivalent or (Description)
Lithological discontinuity	Lithologic discontinuity
Reducing conditions	(Saturation and reduction of iron)
Secondary carbonates	Identifiable secondary carbonates
Stagnic color pattern	(Color pattern indicates saturation and reduction)
Vertic properties	(Presence of slickensides and cracks open 1 cm or more)
Vitric properties	Volcanic glass
Diagnostic Materials	
Artefacts	(Human manufactured material, such as, bricks, glass, pottery, etc.)
Calcaric material	(Strongly effervesces in 1*M* HCl)
Colluvic material	(Sediments from human caused erosion)
Fluvic material	(Recent fluviatile, marine and lacustrine sediments)
Gypsiric material	(Contains 5% or more gypsum)
Limnic material	Limnic materials
Mineral material	Mineral soil material
Organic material	Organic soil material
Ornithogenic material	(Bird excrement)
Sulfidic material	Sulfidic materials
Technic hard rock	(Human-made hard material)
Tephric material	Volcanic glass

Sources: IUSS Working Group WRB 2006, Soil Survey Staff 2006.

Numerical Classification of Soils

The concept of soil properties as a continuum has led to experiments in arranging or ordinations of soil taxa on a numerical basis. Progress in the numerical classification of soils tends to support the potential cited by Whitehead (1925): "Classification is necessary. But unless you can progress from classification to mathematics, your reasoning will not take you very far."

Numerical taxonomy draws on Adansonian principles rather than Linnean principles of classification outlined below (Sneath and Sokal 1962).

Principles of Classification

Adansonian	Linnean
1. Ideal natural taxonomy is one in which taxa have greatest content of information.	1. Distinct kinds of things without intermediates.
2. Every natural feature is of equal weight in constructing a classification.	2. Best method of classifying is to determine "essential nature" of a population and subdivide according to modifications.
3. Affinity is a function of proportion of features in common.	3. Most important characters for classification are those of greatest physiological importance (plant and animal taxonomy especially).
4. Affinity is independent of phylogeny (evolution of the species of genetic theory).	4. Species are basic constituents of higher groups.

Table 6.11. Abbreviated key to reference soil groups in the World Reference Base for soil resources

Abbreviated Key	Soil Group
Soils with a histic or folic horizon 40 cm or more thick	Histosols
Other soils with a hortic, irragric, plaggic, or terric horizon 50 cm or more thick	Anthrosols
Other soils with 20% or more artefacts in upper 100 cm or technic hard rock within 5 cm	Technosols
Other soils having a cryic horizon starting within 100 cm of the surface	Cryosols
Other soils less than 25 cm to continuous hard rock and no calcic, gypsic, or spodic horizon	Leptosols
Other soils with a vertic horizon starting within 100 cm of the surface and periodic cracks most years	Vertisols
Other soils with fluvic soil material within 25 cm and no andic or vertic properties within 100 cm	Fluvisols
Other soils with a natric horizon starting within 100 cm of the surface	Solonetz
Other soils with a salic horizon within 50 cm of the surface	Solonchaks
Other soils with reducing conditions or gleyic color pattern in half or more of volume within 50 cm	Gleysols
Other soils with 30 cm of vitric or andic properties starting within 25 cm of the surface	Andosols
Other soils with a spodic horizon starting within 200 cm of the surface	Podozols
Other soils with a plinthic, petroplinthic, or pisoplinthic horizon starting within a 50 cm depth	Plinthosols
Other soils with a nitric horizon starting within 100 cm of the surface	Nitisols
Other soils with a ferralic horizon starting within 150 cm of the surface	Ferralsols
Other soils with an abrupt textural change within 100 cm and reducing conditions sometime most years	Planosols
Other soils with reducing conditions within a 50 cm depth sometime in most years	Stagnosols
Other soils with a chroma 2 or less mollic horizon 20 cm or more thick and secondary carbonates starting within 50 cm below mollic	Chernozems
Other soils with a mollic horizon and a calcic horizon or secondary carbonate starting within 50 cm below mollic	Kastanozems
Other soils with a mollic horizon	Phaeozems
Other soils with a gypsic or petrogypsic horizon starting within 100 cm of the surface	Gypsisols
Other soils with a duric or petroduric horizon starting within 100 cm of the surface	Durisols
Other soils with a calcic or petrocalcic horizon starting within 100 cm of the surface	Calcisols
Other soils with an argic horizon starting within 100 cm of the surface and albeluvic tonguing	Albeluvisols
Other soils with an argic horizon with CEC_7 of 24 cmol kg^{-1} clay or more starting within 100 cm of the surface and a base saturation (CEC_7) less than 50% in the major part of the 50 to 100 cm depth	Alisols
Other soils with an argic horizon with a CEC_7 less than 24 cmol kg^{-1} clay and a base saturation (CEC_7) less than 50% in the major part of the 50 to 100 cm depth	Acrisols
Other soils with an argic horizon with an apparent CEC_7 of 24 cmol kg^{-1} clay or more starting within 100 mm (200 cm if overlain by loamy sand or coarser) of surface	Luvisols
Other soils with an argic horizon starting within 100 cm (200 cm if overlain by loamy sand or coarser) of surface	Lixisols

continues

Table 6.11. *Concluded.*

Abbreviated Key	Soil Group
Other soils sandy loam or coarser (weighted average) and less than 40 percent gravel to at least 100 cm below surface	Arenosols
Other soils with a cambic horizon starting within 50 cm of surface or an anthraquic, hortic, hydragric, irragric, plaggic, terric, fragic, petroplinthic, pisoplinthic, plinthic, sallic, or vertic horizon starting within 100 cm of surface	Cambisols
Other soils	Regosols

Source: IUSS Working Group WRB 2006.

Hole and Hironaka (1960) built three-dimensional models that approximated the multidimensional relationships among soil properties. The electronic computer has made developments possible that Sneath and Sokal (1962) describe as new or numerical taxonomy and define as “the numerical evaluation of the affinity or similarity between taxonomic units and the ordering of these units into taxa on the basis of their affinities.” Sarkar et al. (1966) used computer analysis to reduce the number of properties selected for a soil classification to a thoroughly correlated few. Arkley (1968) weighted factors and variables according to their communality with other properties, using cumulative communality cluster analysis. He later concluded that cluster analysis is effective with small numbers of distinct soil groups or with small areas, however in large areas with a large number of different soils, clusters either do not exist or are extremely diffuse (Arkley 1976). He also suggested that a coordinate system based upon predefined centroids is the most promising technique to produce an effective numerical soil classification system.

Although numerical taxonomy of soil is in its infancy, the current availability of high-speed computers and increasing interest in quantification of soil data has led to considerable work in this field. When applied to soil classification, ordination methods can reduce original data to a smaller number of variables. These new variables are used as axes in scatter diagrams to observe clusters (Webster 1975, 1990; Gruijter 1977). The resulting clusters however are discontinuous classes that lack coherence (Burrough 1986). One response to these limitations is to use continuous or fuzzy class models (Odeh et al. 1990, 1992).

Some of the disadvantages or problems of numerical taxonomy have been discussed by Simpson (1964). These can be summarized as follows:

1. A single measure of similarity involves an enormous loss of information.
2. Selection, measurement, and coding of multiple characters are highly subjective.
3. Many different kinds of characters must enter into soil classification making it difficult to considered and write all of them into a computer program.

Although techniques are being developed to address the problems of accurately representing the nature of soil properties, the large number of observations required to satisfy the requirements of the models has to date limited the application of numerical classification in soil survey operations.

Perspective

Students should not become overly concerned that so many different soil classification systems are used around the world. Soil classification systems are created to serve the needs of people. Obviously the needs of people are different in various countries, and these needs change with time. Also, the amount of data available differs from one area to another, and classification cannot comfortably require more data than are available. Certainly, international communication in soil science is hindered by the use of many uncorrelated classification systems. Most international publications require that the soils and soil materials used in research be identified by internationally recognized classification of the World Reference Base or *Soil Taxonomy*. However, within individual countries, it is often more important to identify soils by names more easily recognized by the indigenous population and thereby obtain recognition and support for continued research. Although the need to identify the soils in their research by more than one classification places an added responsibility on scientists in these countries, there is little doubt that national classification systems will continue to be used in many parts of the world.

U.S. Soil Taxonomy

7

A soil classification system completely new in design and nomenclature was created and continues to evolve in the United States (Soil Survey Staff 1960, 1964, 1975, 1994, 1999, 2010). The diagnostic horizons and other soil properties used to define taxa within the system were discussed in Chapter 2. The system was developed to serve the National Cooperative Soil Survey of the United States but also envisioned the need to provide a system to include all soils in the world and accommodated new knowledge. Development began about 1950 guided by over 50 years of soil survey within the United States. The first widely circulated version entitled Soil Classification, a Comprehensive System—the 7th Approximation was made available in 1960 at the International Soil Science Society meeting in Madison, Wisconsin (Soil Survey Staff 1960). Although the Soil Survey Staff in the U.S. Department of Agriculture continues responsibility for the system, soil scientists from academic institutions in the United States and numerous other countries have participated and in some cases have led testing of the system and authored significant changes.

Desired Attributes

The following eight attributes were addressed in constructing the system (Soil Survey Staff 1975).

1. Each taxon should have the same meaning to every user. Therefore, specific measurable quantities that can be duplicated by specific measurement techniques should be used as differentiating criteria.
2. The system should be multicategorical. The hierarchy of the categories should have a rationale that allows for first conceiving all soils into broad categories while providing for more detailed separations in lower categories. Ideally each category should be subdivided by no more than about 10 entities at the next lower category because the human mind can simultaneously grasp up to 10 items but has difficulty with more than that number.
3. Defined taxa should be concepts of real soil bodies that exist in nature. The system should provide for all soils known to exist but not provide for all conceivable combinations of soil properties that have not been studied so as not to influence future research. The system should be capable of accepting new observations.

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

4. Differentiae should be soil properties that can be observed in the field or quantitatively measured by reliable techniques. Many soil properties are seasonally dynamic. Criteria need to provide for measurement of these properties by techniques that have a high probability of reproduction. Criteria for which no known technology of measurement is available should not be used.
5. The taxonomy should be capable of modification to accommodate new knowledge with a minimum disturbance to the existing structure of the system. Classification must follow science and be constantly aware that new research will enable more comprehensive examination of the objects classified (soils) as science progresses and more soils are studied in greater detail, often with new analytical technologies.
6. The criteria used should, insofar as possible, keep undisturbed soils and cultivated soil in the same taxa. A primary function of classification is to predict the outcome of human activity on soil. This is best done if uncultivated soils are recognized in the same taxa as their cultivated counterparts. Human manipulation is most intense near the soil surface therefore higher category differentia should most often be selected from among subsoil properties. Drastic alteration of all or nearly all of the soil profile must however result in a different placement of the soil within the system.
7. The taxonomy must provide for all the soil bodies observed within the landscape. Each soil observed must be provided for with a defined range of properties that mutually distinguishes it from adjacent soil.
8. The taxonomy should provide for all known soils and make provision for new knowledge. The originators of *Soil Taxonomy* recognized that development of the system would be uneven, usually favoring those areas where research facilities were more readily available and delayed in remote, little populated areas of the world. This unavoidable bias toward more studied areas of the world, if properly used, would help speed understanding in less studied areas, but the development of the system would be uneven for these very pragmatic reasons.

Students should clearly understand that each taxon exists only as a conceptually developed and quantitatively defined range of soil properties within a continuum of soil properties that are present in real soil bodies. Therefore, if a pedon is sampled, it is a soil that is sampled, not a taxon. Thus, if the pedon sampled has properties that identify it as being of Miami series taxon, it is proper to state, “I sampled a Miami soil” and not proper to state, “I sampled a Miami series” since the entire range defined by the series Miami has not been sampled. Likewise it is proper to state, “The research plots were on Alfisols,” connoting that the research plots were on soils with properties defined by the order Alfisols, not “The research plots were on the Alfisols order.” No single sampling or study site will include the entire range of soil properties defined by a taxon.

Soil properties that can be measured and used as criteria to define taxa are present as continuum among spatially contiguous soils and vertically within soil profiles. Therefore, it is necessary to specify depth and/or thickness limits when defining taxa criteria. Although strict quantitative limits may seem arbitrary, they are needed to factually communicate soil properties via a structured soil classification system.

Unlike many other soil classification systems, soil temperature and soil moisture dynamics so critical to the composition of natural ecosystems, sustained agricultural production and other practical uses of soil, are used as criteria in *Soil Taxonomy*. Currently these criteria utilize long-term climatic data, but direct soil measurements are now being conducted to refine the site-specific conditions that are known to deviate among soils in a climatically defined area because of slope and aspect. Perhaps the most important attribute of the system is provision for changes that accommodate new information without destroying the system. This attribute has been successfully tested many times.

Definition of the Classes

Differentiating characteristics selected are properties of the soils themselves. Concepts and theories of soil genesis are not employed as classification criteria but used as a guide to relevance and weighing of soil properties (Smith 1968). By necessity, methods of analysis must be specified to avoid conflicting data resulting from alternative analytical methods. Definitions are precise and quantitative rather than comparative, and are written in terms that are compatible with current soil survey operations (Smith 1968). As new methods of analyzing soil are developed and tested for consistency and applicability definitions are subject to alteration, new diagnostic criteria may be defined.

Nomenclature

A completely new set of names for taxa above the level of the soil series was considered essential at the time *Soil Taxonomy* was conceived. Historically different concepts and interpretations had been published for the same soil name. Over time different names were often given to the same soil. The sources of names were often from folk terms in different languages as well as coined terms many of which carried no connotation of soil properties. Most named soils were identified only by descriptions of the modal or central concept and no defined range of properties that were mutually exclusive of other named soils were provided. As a result there was much awkwardness in naming intergrades and difficulty in translating in different languages.

Since there is no practical way of updating published names, new names were required. A new nomenclature was devised using mainly classic Greek and Latin sources (Heller 1963). The devised names are connotative insofar as feasible, with the names for the higher categories being relatively short and readily translated into western European languages (Smith 1968; Soil Survey Staff 1975). This is done by successively including a formative element from each of the higher categories to form names in each of the four highest categories. These names may seem awkward and strange at first (indeed, one critical author ascribed an Alice-in-Wonderland aspect to them), but with a little study and experience, one can appreciate the advantages of this connotative nomenclature.

Structure of the System

The system contains six categories. From highest to lowest levels of generalization, these are order, suborder, great group, subgroup, family, and series. The nature of the kinds of differentiating characteristics employed in the various categorical levels of *Soil Taxonomy* is given in Table 7.1.

Two types of comments and explanations should be made to aid in understanding this table and the general operating procedure in selecting differentiating characteristics. First, at the highest categorical level, one considers the entire population of known soils and places them into broad groupings—the soil orders. Once this is accomplished, the practice is then to consider the nature and properties of only the soils within a given order and determine rationale and criteria for defining suborders within only the soils defined by that order. Subsequently, selection of differentiating criteria for great groups considers only soils within each suborder and differentiating subgroup criteria considers only the properties of soils within each great group. Determinations were made of the types of differentia needed to produce the desired suborders, great groups, and lower categories, in an empirical fashion but drawing heavily on the experience of soil survey in the United States.

Some have criticized the fact that differentiating characteristics are not uniformly applied to all soils at a given categorical level. Uniform application is often considered to be optimum or ideal according to classification theory, but in such a diverse and complex population as soils, this is not feasible or practical. Certain types of differentiating characteristics are applicable only to certain taxa.

Table 7.1. Categories of Soil Taxonomy

Category	Nature of Differentiating Characteristics
Order	Presence or absence of major diagnostic horizons, mineralogical properties, and extremes of soil temperature and moisture regime
Suborder	Soil moisture regimes and diagnostic horizons
Great Group	Degree of diagnostic horizon expression within each suborder taxa
Subgroup	Properties that intergrade to taxa in other orders, suborders and great groups, or extragrade to nonsoil material
Family	Particle-size and mineralogy class of control section, soil temperature regime, and cation exchange activity class in most families; calcareous and reaction, depth, rupture-resistance, coating and crack classes in some families
Series	Any consistently identifiable soil property not specifically identified as criteria in a higher category. (Any soil property considered relevant to potential uses can be recognized as a phase of the soil series, i.e., slope, eroded, irrigated, drained, Etc.) (http://www.statlab.iastate.edu/cgi-bin/osd/osdname.cgi)

For example, the degree of soil moisture regimes are excellent and useful differentia at the suborder level of most orders but have no relevance in the Aridisols that are by order definition confined to an aridic soil moisture regime. Hence, the presence or absence of argillic, cambic, calcic, gypsic, or salic horizons, a duripan, or a cryic soil temperature regime are applied to produce useful and relevant suborder taxa within the Aridisols.

An even more striking example is found in the order Histosols (organic soils). Differentiating characteristics used for taxa of levels below the order in mineral soils generally have no relevance in soils with properties dominated by organic material. The degree to which the organic fibers are decomposed carries a large number of accessory and covariant characteristics in organic dominated soils, and content of organic fibers is used as a differentiating characteristic for suborders within the order Histosols. The practice of using different criteria within a taxonomic system is called the *principle of limited applicability of differentiating characteristics*. This means that certain characteristics may reasonably be applied only to portions of the entire population at any given categorical level.

Order Category. Names of the orders, the formative element derived from within the order name used as an identifier at lower levels, derivation or source of the formative element, and the mnemonic or memory device for each are listed in Table 7.2. The formative element of each order is plural, except for the formative element "ox" for Oxisols, which omits the 's' to avoid awkward pronunciation, indicating that there are many soils in that order and is used at the end of all subsequent categories of soils in each order.

Table 7.2. Soil order names and their formative elements

Order	Formative Element	Derivation	Pronunciation Aid[a]
Alfisols	Alfs	Nonsense syllable	Ped*alf*er
Andisols	Ands	Jap. *ando*, "black soil"	A*nd*o
Aridisols	Ids	L. *aridus*, "dry"	Ar*id*
Entisols	Ents	Nonsense syllable	Rec*ent*
Gelisols	Els	L. *gelare*, "to freeze"	J*el*l
Histosols	Ists	Gr. *histos*, "tissue"	H*isto*logy
Inceptisols	Epts	L. *inceptum*, "beginning"	Inc*ept*ion
Mollisols	Olls	L. *mollis*, "soft"	Mo*ll*ify
Oxisols	Ox	F. *oxide*, "oxide"	O*x*ide
Spodosols	Ods	Gr. *spodos*, "wood ash"	O*dd*
Ultisols	Ults	L. *ultimus*, "last"	*Ult*imate
Vertisols	Erts	L. *verto*, "turn"	Inv*ert*

[a]The italicized letters are the formative elements for the order. The pronunciation name associates the pronunciation with a common word. For example, Spodosol (and Othods, Aquods) should be pronounced with a short *o*, as in "odd," not with a long $\bar{o}$, as in "mode."

When attempting to "key out" and identify the name of an unknown soil, it is imperative that a person start at the beginning or top of the key to verify differentiating characteristics for placement. The following simplified key may be used as a general semiquantitative guide to placement of soils in the orders.

Simplified Key to Soil Orders

Criteria	Order
Soils with permafrost or gelic material within 100 cm	Gelisols
Other soils with more than 30% organic matter content to a depth of 40 cm or more	Histosols
Other soils with a spodic horizon within a depth of 200 cm	Spodosols
Other soils with andic soil properties in one-half or more of the upper 60 cm	Andisols
Other soils with an oxic horizon, or containing more than 40% clay in the surface 18 cm and a kandic horizon with less than 10% weatherable minerals	Oxisol
Other soils containing more than 30% clay in all horizons and cracks that open and close periodically	Vertisols
Other soils with some diagnostic subsoil horizon(s) and an aridic soil moisture regime	Aridisols
Other soils with and argillic or kandic horizon and a base saturation percentage at pH 8.2 less than 35 at a depth of 180 cm	Ultisols
Other soils with a mollic epipedon and a base saturation percentage at pH 7 of 50 or more in all depths above 180 cm	Mollisols
Other soils with an argillic, kandic, or natric horizon	Alfisols
Other soils with an umbric, mollic, or plaggen epipedon, or a cambic horizon	Inceptisols
Other soils	Entisols

It is very important to systematically follow this key, and especially the complete one (Soil Survey Staff 1999, 2010), for each pedon. For example, we have noted some students and inexperienced persons, upon finding that a soil has a mollic epipedon, to immediately assume the soil is a Mollisol, which could be incorrect if the pedon fails to have a base saturation percentage of 50 or more in all depths above 180 cm or other Mollisol requirements given in the complete key. Some soils in other orders may have mollic epipedons. Similar keys to all the suborders, great groups and subgroups must be systematically followed to obtain a proper soil name. It is beyond the scope of this text to include complete keys that are available in the following references (Soil Survey Staff 1999, 2010; http:www.statlab.iastate.edu/soils/soiltax).

Suborder Category. The formative elements that identify the suborder taxa, source or derivation of the name, a memory device (mnemonic), and the general meaning of each are listed in Table 7.3. Adding a suborder formative element name prior to the

Table 7.3. Formative elements for suborder names

Formative Element	Derivation	Pronunciation Aid	Meaning or Connotation
Alb	L.-*albus*, white	*alb*ino	Presence of an albic horizon
Anthr	Gr.-*anthropos*, human	*anthr*opology	Specific human modifications
Aqu	L.-*aqua*, water	*aqu*arium	Characteristics associated with wetness
Ar	L.-*arare*, to plow	*ar*able	Mixed horizon
Arg	L.-*argilla*, white clay	*arg*illite	Presence of an argillic horizon
Calc	L.-*calcis*, lime	*calc*ium	Presence of a calcic horizon
Camb	L.-*cambiare*, to exchange	*camb*ist	Presence of a cambic horizon
Cry	Gr.-*kryos*, cold	*cry*olite	Cold
Dur	L.-*duras*, hard	*dur*able	A duripan
Fibr	L.-*fibra*, fiber	*fib*rous	Least decomposed stage
Fluv	L.-*fluvius*, river	*fluv*ial	Flood plains
Fol	L.-*folia*, leaf	*fol*iage	Mass of leaves
Gel	L.-*gelare*, freeze	*jel*lo	Mean annual soil temperature 0°C or less
Gyps	L.-*gypsum*, gypsum	*gyp*sum	Presence of a gypsic horizon
Hem	Gr.-*hemi*, half	*hem*isphere	Intermediate state of decomposition
Hist	Gr.-*histos*, tissue	*hist*ology	Presence of organic soil materials
Hum	L.-*humus*, earth	*hum*us	Presence of organic matter
Orth	Gr.-*orthos*, true	*orth*ophonic	The common ones
Per	L.-*perennis*, all year	*per*ennial	Perudic soil moisture regime
Psamm	Gr.-*psammos*, sand	*psamm*ite	Sand textures
Rend	Polish-*Rend*zina, limestone soil	*Rend*zina	High carbonate content
Sal	L.-*sal*, salt	*sal*ine	Presence of a salic horizon
Sapr	Gr.-*sapros*, rotten	*sapr*ophyte	Most decomposed stage
Torr	L.-*torridus*, hot, dry	*torr*id	Torric soil moisture regime
Turb	L.-*turbidis*, disturbed	*turb*ulence	Presence of cryoturbation
Ud	L.-*udus*, humid	*ud*ometer	Udic soil moisture regime
Ust	L.-*ustus*, burnt	comb*ust*ion	Ustic soil moisture regime
Vitr	L.-*vitrum*, glass	*vitr*ous	Presence of glass
Wass	Ger. *Wasser*, water	*Wasser*mann	Daily under shallow water
Xer	Gr.-*xerox*, dry	*xer*ophyte	Xeric soil moisture regime

Source: Soil Survey Staff 1999, 2010a.

formative element of the order derives the suborder name, such as, Aqualfs are wet Alfisols (*alfs*) preceded by *aqu* connoting wetness.

Great Group Category. Great group formative elements are listed in Table 7.4. Adding a great group formative element name prior to the suborder derives the great group name; for example, Cryaqualfs are cold (*Cry*), wet (*aqu*), Alfisols (*alfs*). Great group, suborder, and order are proper names and always capitalized.

The formative elements for all great groups currently recognized are outlined in Table 7.5. This table can be used to generate the full name of all suborders and great

Table 7.4. Great Group Formative Elements

Formative Element	Derivation	Mnemonicon	Meaning or Connotation
Acro (Acr)[1]	Gr.-*akros*, at the end	*acro*lith	Extreme weathering
Al	Modified from aluminum	*Al*uminum	High aluminum; low iron
Alb	L.-*albus*, white	*alb*ino	Presence of an albic horizon
Anhy	Gr.-*anhydros*, dry	*anhy*drous	Very dry
Anthra	Gr.-*anthropos*, human	*Anthro*pology	Presence of an anthropic epipedon
Aqui (Aqu)[1]	L.-*aqua*, water	*aqu*arium	Wetness
Argi	L.-*argilla*, white clay	*arg*illite	Presence of an argillic horizon
Calci	L.-*calcis*, lime	*calc*ium	Presence of a calcic horizon
Cryo (Cry)[1]	Gr.-*Kryos*, cold	*cry*stal	Cryic soil temperature regime
Duri (Dur)[1]	L.-*duras*, hard	*dur*able	Presence of a duripan
Dystro (Dystr)[1]	Gr.-*dys*, ill; *dystrophic,* infertile	*dystro*phic	Low base saturation
Endo	Gr.-*endon*, within	*endo*carp	Ground water table
Epi	Gr.-*epi*, on, above	*epi*carp	Perched water table
Eutro (Eutr)[1]	Gr.-*eu*, good; fertile	*eutr*ophic	High base saturation
Ferr	L.-*ferrum*, iron	*ferr*ic	Presence of iron
Fibr	L.-*fibra*, fiber	*fibr*ous	Least decomposed stage
Fluv	L.-*fluvus*, river	*fluv*ial	Flood plain
Fol	L.-*folia*, leaf	*fol*iage	Mass of leaves
Fragi	L.-*fragillis*, brittle	*frag*ile	Presence of a fragipan
Fragloss	Compound of *fra(gi)* and *gloss*		See the formative elements *fragi* and *gloss*
Frasi	Ger. *Frasi*, fresh	*Phrase*ology	Not salty
Fulvi (Fulv)[1]	L.-*fulvus*, brownish yellow	*Fulv*ic acid	Dark brown with organic carbon
Glac	L.-*glacialis*, icy	*Glac*ial ice	Presence of ice lenses
Glosso (Gloss)[1]	Gr.-*glossa*, tongue	*gloss*ary	Presence of a glossic horizon
Gypsi	L.-*gypsum,* gypsum	*gyps*um	Presence of a gypsic horizon
Hal	Gr.-*hals*, salt	*hal*ophyte	Salty
Haplo (Hapl; Hap)[1]	Gr.-*haplous*, simple	*hapl*oid	Minimum horizon development
Hemi	Gr.-*hemi*, half	*hem*isphere	Intermediate decomposition
Histo (Hist)[1]	Gr.-*histos*, tissue	*hist*ology	Organic soil material
Humi (Hum)[1]	L.-*humus*, earth	*hum*us	Presence of organic carbon
Hydro (Hydr)[1]	Gr.-*hydro*, water	*hydr*ophobia	Presence of water
Kandi	Modified from kandite	*kan*dite	Presence of a kandic horizon
Kanhaplo (Kanhap)[1]	Compound *kan*(*di*) and *hapl*		Thin kandic horizon
Luvi	Gr.-*louo*, to wash	ab*lu*tion	Illuvial organic material
Melano (Melan)[1]	Gr.-*melasanos*, black	*melan*ic	Presence of melanic epipedon
Molli (Moll)[1]	L.-*mollis*, soft	*moll*ify	Presence of a mollic epipedon
Natri (Natr)[1]	Modified from *natrium*, sodium	*natr*olite	Presence of a natric horizon
Pale	Gr.-*paleos*, old	*pale*osol	Excessive development
Petro (Petr)[1]	Gr.-*petra,* rock	*petro*logy	Petrocalcic horizon
Plac	Gr.-base of *plax*, flat stone	*plac*ard	Presence of a thin pan
Plagg	Ger.-*plaggen*, sod		Presence of a plaggen epipedon

Table 7.4. *Concluded.*

Formative Element	Derivation	Mnemonicon	Meaning or Connotation
Plinth	Gr.-*plinthos,* brick		Presence of plinthite
Psammo (Psamm)[1]	Gr.-*psammos*, sand	*psamm*ite	Sandy texture
Quartzi	Ger.-*quarz,* quartz	quartz	High quartz content
Rhodo (Rhod)[1]	Gr.- *rhodon*, rose	*rhod*odendron	Dark red colors
Sali (Sal)[1]	L.-base of *sal*, salt	*sal*ine	Presence of a salic horizon
Sapr	Gr.-*sapros*, rotten	*sapr*ophyte	Most decomposed stage
Sombri	Fr.-*sombre*, dark	*som*ber	Presence of a sombric horizon
Sphagno	Gr.-*sphagnos*, bog	*sphag*num moss	Presence of sphagnum moss
Sulfo (Sulf; Sulfi)[1]	L.-*sulfur*, sulfur	*sulf*ur	Presence of sulfides or their oxidation products
Torri	L.-*torridus,* hot and dry	*torr*id	Torric (aridic) soil moisture regime
Udi (Ud)[1]	L.-*udus*, humid	*ud*ometer	Udic soil moisture regime
Umbri (Umbr)[1]	L.-base of *umbra*, shade	*umbr*ella	Presence of an umbric epipedon
Usti (Ust)[1]	L.-base of *ustus*, burnt	comb*ust*ion	Ustic soil moisture regime
Verm	L.-base of *vermes*, worm	*verm*iform	Wormy, or mixed by animals
Vitri (Vitr)[1]	L.-*vitrum*, glass	*vitr*eous	Presence of glass
Xero (Xer)[1]	Gr.-*xerox*, dry	*xero*phyte	Xeric soil moisture regime

Source: Soil Survey Staff, 1999, 2010a.
[1] (Alternative spelling used for some Great Groups).

groups. As an example, if the order name of a soil is Alfisols, the great group name ends with the order formative element "alfs." If the soil has an aquic soil moisture regime, the suborder formative element "aqu" precedes the order formative element forming the suborder name Aqualfs. If the soil has a fragipan, identified as "Fragi" in the "alfs" row and in the "aqu" suborder column, the great group name is Fragiaqualfs. The row listing of orders and the column listing of suborder formative elements are alphabetical. The listing of great group formative elements under each suborder column in each order row is sequential as the great groups are keyed in each suborder (Soil Survey Staff 2010). Also note that the spelling of some great group formative elements differs slightly among different suborders. Differences usually relate to adding or deleting a suffix vowel to the formative element. This is done to facilitate pronunciation of the great group name. For example, in the Alfisols 'Natr' pronounces well when followed by a vowel as in the Aqualfs suborder, that is, Natraqualfs. However, when used in the Xeralfs suborder, 'Natra' is used for the great group Natraxeralfs. This table is a useful checklist of all suborders and great groups (Soil Survey Staff 2010).

Subgroup Category. Subgroups are identified with a binomial nomenclature. Subgroup names include the great group name, as a capitalized noun, preceded by a capitalized modifying subgroup adjective(s) name identifying the main differentiating

Table 7.5. Great group formative elements arranged according to orders (rows) and suborders (columns) with great groups listed in suborder columns according to suborder keys

Order Formative Elements	Suborder Formative Elements													
	alb	anthr	aqu	ar	arg	calc	camb	cry	dur	fibr	fluv	fol	gel	gyps
alfs (Alfisols)			Cry Plinth Dur Natr Fragi Kandi Verm Alb Gloss Epi Endo					Pale Glosso Haplo						
ands (Andisols)			Gel Cry Plac Dur Vitr Melan Epi Endo					Duri Hydro Melano Fulvi Vitri Haplo					Vitri	
ids (Aridisols)					Petro Natr Pale Gypsi Calci Hapl	Petro Haplo	Aqui Petro Anthra Haplo	Sali Petro Gypsi Argi Calci Haplo	Natri Argi Haplo					Petro Natri Argi Calci Haplo
ents (Entisols)			Sulf Hydr Gel Cry Psamm Fluv Epi Endo	Ust Xer Torri Ud							Geli Cryo Xero Usti Torri Udi			
els (Gelisols)														
ists (Histosols)										Cryo Sphagno Haplo		Cryo Torri Usti Udi		
epts (Inceptisols)		Plagg Hapl	Sulf Petr Hal Fragi Gel Cry Verm Hum Epi Endo					Humi Calci Dystro Haplo					Humi Dystro Haplo	

Suborder Formative Elements (continued)

hem	hum	ist*	orth	per	psamm	rend	sal	sapr	torr	turb	ud	ust	vitr	wass	xer
											Natr Ferr Fragloss Fragi Kandi Kanhapl Pale Rhod Gloss Hapl	Dur Plinth Natr Kandi Kanhapl Pale Rhod Hapl			Dur Natri Fragi Plintho Rhodo Pale Haplo
									Duri Vitri Haplo		Plac Dur Melan Hydr Fluv Hapl	Dur Hapl	Usti Udi		Vitri Melano Haplo
							Aqui Haplo								
			Gel Cry Torri Xer Ust Ud		Cryo Torri Quartzi Usti Xero Udi									Frasi Psammo Sulfi Hydro Fulvi Haplo	
		Fol Glac Fibr Hem Sapr	Hist Aqu Anhy Moll Umbr Agri Psamm Hapl							Histo Aqui Anhy Molli Umbri Psammo Haplo					
Sulfo Sulfi Luvi Cryo Haplo								Sulfo Sulfi Cryo Haplo						Frasi Sulfi Haplo	
											Sulf Dur Fragi Hum Eutr Dystr	Dur Calci Hum Dystr Hapl			Duri Fragi Humi Calci Dystro Haplo

continues

Table 7.5. *Concluded.*

Order Formative Elements	Suborder Formative Elements													
	alb	anthr	aqu	ar	arg	calc	camb	cry	dur	fibr	fluv	fol	gel	gyps
olls (Mollisols)	Natr Argi		Cry Dur Natr Calci Argi Epi Endo					Dur Natri Pale Agri Calci Haplo					Haplo	
ox (Oxisols)			Acr Plinth Eutr Hapl											
ods (Spodosols)			Cry Al Fragi Plac Dur Epi Endo					Placo Duri Humi Haplo					Humi Haplo	
ults (Ultisols)			Plinth Fragi Alb Kandi Kanhapl Pale Umbr Epi Endo											
erts (Vertisols)			Sulf Sal Dur Natr Calci Dystr Epi Endo					Humi Haplo						

characteristic of that subgroup. For example, the subgroup name Humic precedes the great group name Fragiaqualfs forming the complete subgroup name Humic Fragiaqualfs.

Within each great group, four types of subgroups can be formed:

1. "Central concept" subgroups have properties most representative of the great group and are named "Typic" (*typical*) or "Haplic" (*simple*). All great groups have a Typic or Haplic subgroup, such as, Typic Argiudolls, Haplic Xerarents, etc. The Typic or Haplic subgroups are sequentially the last subgroup within each great group key and usually identified only as "other (name of great group)."
2. "Intergrade" subgroups have specific properties that differ from the Typic or Haplic subgroup and have one or more characteristics similar to another order,

Suborder Formative Elements (continued)

hem	hum	ist*	orth	per	psamm	rend	sal	sapr	torr	turb	ud	ust	vitr	wass	xer
						Cry Hap					Natr Calci Pale Argi Verm Hapl	Dur Natr Calci Pale Agri Verm Hapl			Duri Natri Pale Calci Agri Haplo
				Sombri Acro Eutro Kandi Haplo					Acro Eutro Haplo		Sombri Acr Eutr Kandi Hapl	Sombri Acr Eutr Kandi Hapl			
	Placo Duri Fragi Haplo		Plac Dur Fragi Al Hapl												
	Sombri Plintho Kandi Kanhaplo Pale Haplo										Plinth Fragi Kandi Kanhapl Pale Rhod Hapl	Plinth Kandi Kanhapl Pale Rhod Hapl			Pale Haplo
									Sali Gypsi Calci Haplo		Dystr Hapl	Dystr Sal Gypsi Calci Hapl			Duri Calci Haplo

suborder, or great group. The subgroup name is the adjective form of the taxon they resemble but are defined by somewhat different properties and are formed by adding “ic” to the singular form of the taxon name. For example, Vertic Hapludolls are Hapludolls with certain properties that are common to Vertisols, but these properties are not well enough expressed to classify the soil in the Vertisol order. Similarly, Fluventic Hapludolls have some properties of soils belonging to the suborder Fluvents (flooding, Entisols), and Fluvaquentic Hapludolls have some properties of soils belonging to the great group Fluvaquents (flooding, aquic, Entisols) although their major properties place them in the Hapludolls great group of Mollisols.

3. “Extragrade” subgroups identify soil individuals with properties that do not clearly intergrade toward specifically defined categories but have one or more

properties common to soils in several categories or to nonsoil material. Lithic subgroups, indicating a shallow depth to nonsoil bedrock, and Calcic subgroups, indicating the presence of a calcic horizon that is present in several kinds of soil, are examples of extragrade subgroups.

4. "Intragrade" subgroups identify specific properties that differ from the Typic subgroup of that great group. For example, Xanthic Hapludox have a more yellow color than the Typic Hapludox, and Aeric Argiaquolls are somewhat better drained than Typic Argiaquolls. Extragrade and intragrade subgroup formative elements are listed in Table 7.6.

Some subgroups use more than one name such as Lithic Ultic Argixerolls indicating Argixerolls that are shallow to a lithic contact and chemical properties approaching those of Ultisols.

Family Category. Family taxa serve both to further subdivide subgroups and aggregate soils differentiated by higher categories within *Soil Taxonomy*. The family category was intended to be useful for making major interpretations for growing plants and engineering purposes (Smith 1986). Criteria selected to define most families are soil properties not easily modified by available human technology regardless of associated soil properties identified by higher category classes. Where higher categories have utilized some of these criteria, they are not used again in the family category thereby avoiding redundancy in the system. Certain soil properties that present unique constraints to use and management are present in only a few soils, and some family classes are used only for specific soils as identified by higher categories.

Family category classes follow:

- Particle-size classes
- Substitute particle-size classes
- Mineralogy classes
- Cation-exchange activity classes
- Calcareous and reaction classes
- Soil temperature classes
- Soil depth classes
- Rupture-resistance classes
- Classes of coatings
- Classes of cracks

Particle-Size Classes. Family particle-size classes are determined in the particle-size control section of each soil. For most soils, the particle-size control section is that volume of a pedon between 25- and 100-cm deep or in the root-limiting layer if shallower. If a root-limiting layer occurs at less than 36 cm, the control section extends from the surface to that depth. If a pedon has an argillic, kandic, or natric horizon,

Table 7.6. Subgroup formative elements that extragrade to properties not identified by higher taxa or intragrade[a] within a great group

Formative Element	Derivation	Connotation
Abruptic	L.-*abruptum*, torn off	Abrupt textural change
Acric[a]	Gr.-*akros,* at the end	Low apparent CEC
Aeric[a]	Gr.-*aerios*, air	More aeration than Typic subgroup
Albic	L.-*albus*, white	Presence of albic materials
Albaquic		(see Albic and Aquic)
Alic	L.-*alumen*, alumina	High extractable Al content
Anionic	Gr.-*anion*, neutral	Low CEC or positively charged
Anthraquic	Gr.-*anthropos*, human + L.-*aqua*, water	Human controlled flooding as in paddy rice culture
Anthropic	Gr.-*anthropos*, human	An anthropic epipedon
Aquic[a]	L.-*aqua*, water	Wetter than Typic subgroup
Arenic[a]	L.-*arean*, sand	50 to 100 cm sandy textured surface
Argic[a]	L.-*argilla,* clay	Argillic horizon
Calcic	L.-*calcis*, lime	Presence of a calcic horizon
Chromic[a]	Gr.-*chroma*, color	High chroma colors
Cumulic[a]	L.-*cumulus*, heap	Thickened epipedon
Duric	L.-*durus*, hard	Presence of a duripan
Durinodic	L.-*durus* + L.-*nodus*, hard knot	Presence of durinodes
Dystric[a]	Gr.-*dys*, ill	Lower base saturation percentage
Eutric[a]	Gr.-*eu*, good, fertile	Higher base saturation percentage
Fibric	L.-*fibra*, fiber	>25 cm of fibric material
Fragiaquic		(see Fragic and Aquic)
Fragic	L.-*fragilis*, brittle	Presence of fragic properties
Glacic	L.-*glacialis*, icy	Presence of ice lenses or wedges
Glossaquic		(see Glossic and Aquic)
Glossic	Gr.-*glossa*, tongue	Interfingered horizon boundaries
Grossarenic[a]	L.-*grossus*, thick and L.- *arena*, sand	>100 cm sandy textured surface
Gypsic	L.-*gypsum*, gypsum	Presence of a gypsic horizon
Halic	Gr.-*hols*, salt	Salty
Hemic	Gr.-*hemi*, half	>25 cm of hemic organic material
Humic[a]	L.-*humus*, earth	Higher organic matter content
Hydric	Gr.-*hydor*, water	Presence of water
Kandic	Modified from Kandite	Low activity clay present
Lamellic	L.-*lamella*, dim	Presence of lamellae
Leptic[a]	Gr.-*leptos*, thin	Thinner than Typic subgroup
Limnic	Gr.-*limne*, lake	Presence of a limnic layer
Lithic	Gr.-*lithos*, stone	Shallow lithic contact present
Natric	Modified from *natrium*, sodium	Presence of sodium
Nitric	Modified from *nitron*	Presence of nitrate salts
Ombroaquic[a]	Gr.-*ombros*, rain + L.-*aqua*	Surface wetness
Oxyaquic[a]	F.-*oxygen* + L.-*aqua*	Water saturated but not reduced
Pachic[a]	Gr.-*pachys*, thick	Epipedon thicker than Typic subgroup

continues

Table 7.6. *Concluded.*

Formative Element	Derivation	Connotation
Petrocalcic	Gr.-*petra*, rock and calcic	Presence of a petrocalcic horizon
Petroferric	Gr.-*petra*, rock and L.-*ferrum*, iron	Presence of petroferric contact
Petrogypsic	Gr.-*petra* + L.-*gypsum*	Presence of a petrogypsic horizon
Petronodic	Gr.-*petra*, + L.-*nodus*, rock knots	Presence of concretions and/or nodules
Placic	Gr.-*plax*, flat stone	Presence of a placic horizon
Plinthic	Gr.-*plinthos*, brick	Presence of plinthite
Plinthaquic		(see Plinthic and Aquic)
Rhodic[a]	Gr.-*rhodon*, rose	Darker red colors than Typic subgroup
Ruptic	L.-*ruptum*, broken	Intermittent horizon
Salic	L.-*sal*, salt	Presence of a salic horizon
Sapric	Gr.-*sapros*, rotten	>25 cm of sapric organic material
Sodic	F-*soda*, sodium	High exchangeable Na content
Sombric	F.-*sombre*, dark	Presence of a sombric horizon
Sphagnic	Gr.-*sphagnos*, a moss	Sphagnum organic material
Sulfic	L.-*sulfur*, sulfur	Presence of sulfides
Terric[a]	L.-*terra*, earth	Mineral substratum within 1 meter
Thaptic (Thapto-)	Gr.-*thapto*, buried	A buried soil horizon present
Turbic[a]	L.-*turbibis,* disturbed	Evidence of cryoturbation
Udic[a]	L.-*udis*, humid	More humid than Typic subgroup
Umbric	L.-*umbra*, shade	Presence of an umbric epipedon
Ustic[a]	L.-*ustus*, burnt	More ustic than Typic subgroup
Vermic	L.-*vermes*, worms	Animal mixed material
Vitric	L.-*vitrum*, glass	Presence of glassy material
Xanthic[a]	Gr.-*xanthos*, yellow	More yellow colors than Typic subgroup
Xeric[a]	Gr.-*xerox*, dry	More xeric than Typic subgroup

[a]Intragrades that indicate a deviation of a soil property from the Typic or Haplic subgroup of that great group.

with an upper boundary less than 100-cm deep, the upper 50 cm of that horizon, or the entire horizon if less than 50-cm thick, is the control section.

All mineral soils, except those with andic soil properties or a high content of volcanic glass, cinders, or pumice, receive a particle-size family name. Terric subgroups of Histosols and Histels also have particle-size families based on the particle-size class of the mineral layer. Most mineral soils are classified by the particle-size classes outlined in Figure 2.2B, Chapter 2. Note that the family particle-size classes utilize the engineering definition of silt as particles of 2–74 μm in diameter rather than 2–50 μm in diameter used in traditional particle-size classification of soil. Also, in Figure 2.2B soil material with more than 35% clay is referred to as clayey but subdivided at 60% clay into fine, 35 to less than 60% clay and very fine, 60% of more clay subclasses. The subclasses fine and very fine are used for all mineral soils, but the term clayey is used for Histosols and Histels and for material that contains more the 35% rock fragments.

If the control section volume of any soil is 90% or more rock fragments and voids and contains less than 10% particles less than 2.0 mm in diameter, the family class is termed *Fragmental*. All particle-size classes in Figure 2.2B may be modified as *skeletal*, if the control section contains more than 35% of rock fragments greater than 2 mm in diameter. Only *sandy-skeletal*, *clayey-skeletal* and *loamy-skeletal* designations are made, both fine-loamy and coarse-loamy classes being grouped as *loamy* and fine and very-fine classes grouped as *clayey*, if more than 35% rock fragments are present.

In some soils the control section may have contrasting particle-size classes. If there is a substantial and clear (within a vertical distance of 12.5 cm) change in particle-size within the control section, contrasting particle-size classes are used to identify the family. Examples are *fine-loamy over sandy or sandy-skeletal*; *hydrous over clayey*; *sandy over clayey,* etc. Sixty-four such strongly contrasting series have been designated. (See Soil Survey Staff 2010 for a complete list.) If the control section includes more than one pair of strongly contrasting classes, the family is named for the most contrasting, adjacent, particle-size classes followed by the class name *Aniso* (from the Greek, *Anisos,* unequal).

In some soils, the particle-size of the control section has been used as differentia in higher categories. Psamments are defined as being sandy, thus no particle-size family is recognized to avoid redundancy. Vertisols must contain more than 30% clay. To avoid a very small number of pedons having between 30 and 35% clay in the control section, Vertisols with 30–60% clay in the control section are classified as *fine* in the family category. Shallow soils, Lithic, Terric, Arenic, Grossarenic, and Pergellic subgroups, are identified only as *loamy,* for both fine-loamy and coarse-loamy control sections, *silty* for both fine-silty and coarse-silty, and *clayey* for fine and very-fine classes. These classes are modified as *skeletal* (for example, *clayey-skeletal*, etc.) if they contain more than 35% coarse fragments.

Substitute Particle-size Classes. The particle-size classes do not adequately classify soil materials with a volumetric fine earth (particles less than 2 mm) content less than 10%, or more 60% (by weight) cinders, pumice, or volcanic glass, and materials with andic properties (as shown in Figure 2.4) in the thickest part of their particle-size control section. Andic materials do not disperse well upon particle-size analysis, and substitute classes identified by water retention and rock type are used. In place of the more ordinary particle-size classes, the substitute particle-size classes in Table 7.7 are used in the family category.

Mineralogy Classes. Family mineralogical classes are based on the mineral composition of the particle-size control section. The mineral composition of many soils is limited by criteria of the higher categories. Therefore, mineral class criteria to identify certain mineral class families may differ slightly among soils

Table 7.7. Key to suzbstitute particle-size family classes[1]

Approximate Control Section Criteria	Family Particle-Size Class
More than 60%[2] volcanic ash, cinders, lapilli, pumice and pumice-like particles, >67% of which is pumice and pumice-like	Pumiceous
More than 60% volcanic ash, cinders, lapilli, pumice and pumice-like particles, <67% of which is pumice and pumice-like	Cindery
Less than 10% of the volume is fine earth (<2 mm) material	Fragmental
AndicX soil properties and water content at 1,500 kPa <12% (dried); <30% (undried) or 30% or more of the >2 mm particles are glass, glass-coated grains, or vitric volcaniclastics, and:	
35% or more rock fragments >67% of which are pumice or pumice-like	Ashy-pumiceous
35% or more rock fragments	Ashy-skeletal
Less than 35% rock fragments	Ashy
Andic soil properties and water content at 1,500 kPa >12% (dried) or 30–100% (undried) and:	
35% or more rock fragments >67% of which are pumice or pumice-like	Medial-pumiceous
35% or more rock fragments	Medial-skeletal
Less than 35% rock fragments	Medial
Andic soil properties and water content at 1,500 kPa of 100% or more (undried) and:	
35% or more rock fragments >67% of which are pumice or pumice-like	Hydrous-pumiceous
35% or more rock fragments	Hydrous-skeletal
Less than 35% rock fragments	Hydrous
Soils with 40% or more gypsum <20 mm in diameter and:	
35% or more rock fragments	Gypseous-skeletal
Less than 35% rock fragments and 50% or more particles with diameters of 0.1 to 2.0.	Coarse-gypseous
Less than 35% rock fragments	Fine-gypseous
For all other mineral soils and Terric subgroups of Histosols and Histels, see Figure 2.2B.	

[1] Use key systematically.
[2] Percentages are based on volumetric measurements except coarse-gypseous definition is weight based.

in different higher taxa. Terric subgroups of Histosols and Histels are assigned mineralogy classes as determined in the mineral section of the pedon. Quartzipsamments, being restricted to more than 90% quartz, which is the same as siliceous differentia at the family category, use neither particle-size nor mineralogy family names. Table 7.8 is a key to appropriate mineralogy classes within higher category groupings.

Cation Exchange Activity Classes. Cation exchange activity class is the apparent cation exchange capacity (CEC_7/% clay) of the less than 2 mm fraction in the control

Table 7.8. Key to family mineralogy classes[1]

Approximate Control Section Criteria	Family Mineralogy Class
For Oxisols and "kandi" and "kanhap" great groups of Ultisols and Alfisols	
More than 40% Fe_2O_3 (>28% Fe) in <2 mm fraction	Ferritic
More than 40% gibbsite in <2 mm fraction	Gibbsitic
Between 18 and 40% of both gibbsite and Fe_2O_3 in <2 mm fraction	Sesquic
18 to 40% Fe_2O_3 in <2mm fraction	Ferruginous
18 to 40% gibbsite in <2 mm fraction	Allitic
More than 50% kaolinite; other 1:1 minerals; gibbsite, non-expanding 2:1 minerals in <2 mm clay and less halloysite than kaolinite	Kaolinitic
More than 50% halloysite in <2 mm clay	Halloysitic
Other soils in this group	Mixed
For Ashy, Medial, Hydrous, Cindery, and Other Substitute Particle-size classes other than Fragmental	
Have 40% or more (by weight) gypsum in either fine earth or <20 mm fraction.	Hypergypsic
Sum of 8 times the Si extracted with ammonium oxalate plus 2 times the Fe extracted with ammonium oxalate of 5 or more, and 8 times Si >2 times Fe	Amorphic
Sum of 8 times the Si extracted with ammonium oxalate plus 2 times the Fe extracted with ammonium oxalate of 5 or more	Ferrihydritic
30% or more volcanic glass in 0.02 to 2.0 mm fraction	Glassy
All other Ashy, Medial, Hydrous, Cindery, and Pumiceous Particle-size Families	Mixed
For all other mineral soils and Terric subgroups of Histosols and Histels	
15% or more gypsum in either <20 mm or <2 mm size fraction	Gypsic
More than 40% carbonates in either <20 mm or <2 mm size fraction	Carbonatic
More than 40% Fe_2O_3 (dithionite citrate extractable) in <2 mm size fraction of nonfragmental families	Ferritic
More than 40% gibbsite and boehmite in <2 mm size fraction of nonfragmental families	Gibbsitic
More than 40% magnesium-silicate minerals in <2 mm size fraction of nonfragmental families	Magnesic
More than 20% glauconitic pellets in <2 mm fraction	Glauconitic
For all other clayey, clayey-skeletal, fine, or very-fine classes of mineral soils and Terric subgroups of Histosols and Histels	
More than 10% gibbsite plus Fe_2O_3 (citrate-dithionite) in <2 mm size fraction of nonfragmental families	Parasesquic
More than 50% of the <0.002 mm fraction is halloysite, kaolinite, and allophane with more halloysite than any other mineral	Halloysitic
More than 50% kaolinite and other 1:1 minerals and less than 10% smectite in <0.002 mm fraction	Kaolinitic
More smectite (montmorillonite, beidellite, and nontronite) than any other mineral in <0.002 mm fraction	Smectitic
More than 50% illite (>4% K_2O) in <0.002 mm fraction	Illitic
More vermiculite than any other clay mineral in <0.002 mm fraction	Vermiculitic

continues

Table 7.8. *Concluded.*

Approximate Control Section Criteria	Family Mineralogy Class
No free carbonates; pH in 50 ml of 1 M NaF >8.4 in 2 minutes; ratio of 1,500 kPa water to measured clay is 0.6 or more	Isotic
Other soils in this category	Mixed
All other mineral soils, except Quartzipsamments	
More than 45% mica and stable mica pseudomorphs in 0.02 to 0.25 mm fraction	Micaceous
Total weight of Fe_2O_3 (citrate-dithionite) plus gibbsite more than 10%	Parasesquic
No free carbonates; pH in 50 ml of 1 M NaF >8.4 in 2 minutes; ratio of 1,500 kPa water to measured clay is 0.6 or more	Isotic
More than 90% quartz chalcedony or opal in the 0.02 to 2.0 mm fraction	Siliceous
All other properties	Mixed
All Histosols and Histels except Terric subgroups	
Have ferrihumic material in control section	Ferrihumic
5 cm or more of coprogenous earth	Coprogenous
5 cm or more of diatomaceous earth	Diatomaceous
5 cm or more of marl	Marly

[1]Use key systematically.

section. Cation exchange activity classes are used for all mineral soils except Oxisols, "Kandi," and "Kanhap" great groups of Alfisols and Ultisols, where they would be redundant and are not used in sandy, sandy-skeletal, or fragmental particle-size families. In contrasting particle-size classes, the material in the control section with the greatest clay content is used. If the ratio of percent water retained at 1,500 kPa, tension to percentage of clay determined by particle-size analysis is 0.25 or less or 0.6 or more, the percent clay used in the calculation is estimated as 2.5 times the value of weight percent water retained at 1,500 kPa tension minus percent organic carbon. The cation exchange activity classes are given in Table 7.9.

Calcareous and Reaction Classes. The presence of calcareous material and reaction (pH value) of a soil are most often identified by criteria used in higher categories of classification. However, in some categories, it has been found necessary to identify these properties at the family level.

An *allic* reaction class is used only in Oxisols that have a layer more than 30-cm thick in the control section that contains more than 2 cmol(+) of KCl-extractable Al per Kg of less than 2 mm material. In these low CEC soils, these criteria identify soils that have higher lime requirements than most other Oxisols.

Acid, nonacid, and calcareous reaction classes are used for many nonsandy Entisols, Gelisols, and Aquands (except Duraquands and Placaquands), and all Aquepts

Table 7.9. Cation-Exchange Activity Classes

Criteria (CEC_7/% clay)	Cation-Exchange Activity Class
0.60 or more	Superactive
0.40 to 0.60	Active
0.24 to 0.40	Semiactive
Less than 0.24	Subactive

except Sulfaquepts, Petraquepts and Fragiaquepts with less than pH 5.0 in 0.01 M $CaCl_2$ (5.5 in 1:1 H_2O) throughout the loamy or finer particle-size control section.

A *nonacid* reaction class is used if pH values are greater than those of the acid reaction class in soils where acid reaction is used.

A *Calcareous* reaction class is used when all parts of the control section or a 2.5-cm layer in soils with a root-restricting layer within 50 cm of the surface effervesce when cold, dilute HCl is applied. In addition to the soils already listed above, *calcareous* is used in Aquolls except Calciaquolls, Natraquolls, and Argiaquolls, or Cryaquolls and Duraquolls that have an argillic horizon. It is not used if the soil qualifies as a *carbonatic, gypsic, or hypergypsic* mineralogy family.

Soil Temperature Classes. Soil temperature regime classes are used in all soils except Cryic ("cry") suborders, great groups, and subgroups. Cryic soils have no permafrost but a mean annual soil temperature less than 8°C. Cryic soils not saturated with water any time during the summer also have mean soil temperature of the three warmest months (summer) less than 15°C if no O horizon is present, or less than 8°C if an O horizon is present. If the soil is saturated at some time during the summer, the mean three-month summer soil temperature is less than 13°C if an O horizon or histic epipedon is present, and the mean summer temperature is less than 6°C. Therefore, cryic soils have a frigid soil temperature class, and it would be redundant to assign a family soil temperature class. Soil temperature classes are outlined in Table 7.10.

Soil Depth Classes. Soil depth classes are assigned in all mineral soils, except Lithic subgroups that have a root-limiting layer present. Root-limiting layers are duripans, petrocalcic, petrogypsic, and placic horizons, continuous ortstein (90% or more), densic, lithic, paralithic, and petroferric contacts.

Shallow families are used to identify soils where the upper boundary of a root-limiting layer is within 50 cm of the surface or within 100 cm of the surface in Oxisols.

Soil depth classes in Histosols recognize the upper boundary of the same root-limiting features as in the mineral soils and in addition the upper boundary of cindery, fragmental, or pumiceous material. The *Micro* depth class identifies a

Table 7.10. Family soil temperature classes

Criteria (Mean Annual Soil Temperature)	Class
Gelisols and Gelic suborders and great groups	
−10°C or lower	Hypergelic
−4°C to −10°C	Pergelic
+1°C to −4°C	Subgelic
Soils that have a mean June, July, and August temperature that differs from the mean December, January, and February temperature by 6°C or more	
Less than 8°C and not Cryic[a]	Frigid
8°C to 15°C	Mesic
15°C to 22°C	Thermic
22°C or higher	Hyperthermic
Soils that have a mean June, July, and August temperature that differs from the mean December, January, and February temperature by less than 6°C	
Less than 8°C and not Cryic[a]	Isofrigid
8°C to 15°C	Isomesic
15°C to 22°C	Isothermic
22°C or higher	Isohyperthermic

[a]Mean soil temperature of warmest three months lower than 15°C. (See Chapter 2.)

root-limiting contact at a depth of less than 18 cm from the surface. The *Shallow* depth class applies if the root-limiting surface is between 18 and 50 cm of the surface. Only the *micro* class is used in Folists.

Rupture-Resistant Class. Ortstein is the only rupture-resistant class and used only in Spodosols. Ortstein families have a partially cemented spodic horizon.

Classes of Coatings. Classes of coating are use only in the Quartzipsamments great group. Coated classes are recognized if in the particle size control section the soil contains more than 5% silt plus clay.
Uncoated classes are used if the control section has less than 5% silt plus clay.

Classes of Cracks. Classes of cracks are used only in great groups of Fluvaquents and Humaquepts.

Cracked families are named when the soil has a layer 50-cm or more thick with permanent lateral and vertical cracks, 2 mm or wider, spaced at lateral intervals of less than 50 cm. These permanent cracks are known to persist for several hundred years and allow water to rapidly pass through the soil without wetting the matrix, for example, bypass flow.

Forming Family Names. The complete family name of a soil is formed by arranging the applicable family classes in the following order: particle-size, mineralogy, reaction, activity, (other), temperature, and depth, followed by the subgroup name, such as fine-loamy, mixed, active thermic, Typic Hapludalfs. Classes that are not applicable to the soil are not included in the family name.

Soil families further separate soils defined by subgroup categories but also provide a mechanism to associate soils separated by differentia used at higher categories. For example, if soil temperature is a major consideration for a specific soil interpretation, it may be most convenient and appropriate to refer to all soils in frigid families for that interpretation. When particle size and mineralogy are major factors relating to a given interpretation, all soils of fine and very-fine, smectitic, or fine. Kaolinitic families, for example, may be an appropriate designation to identify a group of soils for that particular interpretation.

Soil Series Category. A soil series is the most homogeneous grouping of soils made in *Soil Taxonomy*. Each series must have properties that conform to all of the criteria of higher categories. However, in addition to criteria used at higher categories, series criteria can be selected from a control section that extends from the soil surface to and extending to 25 cm below a densic or paralithic contact if the upper boundary of the contact is less than 125 cm below the soil surface. Also, properties below the family particle size control section but above 200 cm can be considered for series criteria. In practice this encourages soil scientists that encounter soils that are classified the same according to all the criteria, order through family, to propose new series within families when they can demonstrate that a new series permits the following: (1) better and more concise recommendations for soil use, (2) that the soil properties proposed as series criteria can be consistently identified, and (3) that soils defined by the proposed new series are present in a large enough area(s) to justify the effort needed to formally develop and publish information related to the new series.

Soil series names, without reference to higher categories, do not connote soil properties. Therefore, in scientific publications, complete taxonomic identification should be provided. For example, soils of the Mohave series are identified as Mohave soils (fine-loamy, mixed, superactive, thermic, Typic Calciargids).

History of Soil Series. The concept of soil series has changed over time creating confusion among those who attempt to identify soils from series names in older literature. For this reason a brief history of the concepts of soil series as used in the United States is presented.

At the inception of soil survey in the United States, three categories were employed in classifying soils in the field: series, type, and phase. Series names are place names of cities, towns, etc., from the area where the soil was first defined. Type identified the texture of the surface horizon, and phase referred to slope, rockiness, and other features of the area being identified. The grouping of soils into categories above series was not discussed in the *Soil Survey Manual* of 1937 (Kellogg 1937). At that time, a series was defined as "a group of soils having genetic horizons similar as to differentiating characteristics and arrangement in the soil profile, and developed from a particular type of parent material" (Kellogg 1937, p. 88). It was also stated that, "A further restriction on the observable differentiation of soil characteristics is that such characteristics must be mappable" and "All mappable differences in the soil profile significant to the growth of plants should be recognized in the classification" (Kellogg 1937, p. 89). Clearly spatial area that could be identified on a soil map was series criteria in this early concept, and the phrase "having genetic horizons similar as to differentiating characteristics" allowed individual interpretations as to the meaning of similar.

The concept of "mappable differences" as criteria for defining soil series resulted in considerable confusion as attempts were made to identify an individual unit of soil that could be classified. Mappable differences in one area within which soils belonging to a series because of profile similarities did not prove to be "mappable" in other areas or at different mapping scales. By 1951 the problem of aligning "genetic horizon" criteria with "mappability" was recognized, but no clear protocol was established to overcome the problem (Soil Survey Staff 1951). To clarify the use of soil series as a category in soil classification, the *7th Approximation* (Soil Survey Staff 1960, p. 15) defined soil series as "a collection of soil individuals essentially uniform in differentiating characteristics and arrangement of horizons; or, if genetic horizons are thin or absent, a collection of soil individuals that, within defined depth limits, are uniform in all soil properties diagnostic for series." "Soil individuals are real things, but series are conceptual." Three contexts were identified for using a soil series name:

1. A series name can be used to identify a soil profile as one individual belonging to that named series. For example, "this profile is an individual in the (named) series."
2. A series name can be used as part of a map unit name to identify the soil individuals delineated by that map unit. For example, "this (named series) map unit contains mainly soil individuals of the (named) series and individuals of (other named series)."
3. A series name can be used as a taxonomic class of soil with a defined range of properties.

By 1993, soil series were defined "as a class, a series is a group of soils or polypedons that have horizons similar in arrangement and in differentiating characteristics" (Soil Survey Staff 1993, p. 20). Mappability as a criterion was dropped from the concept of a series, and it was emphatically stated that, "when the limits of soil taxa are superimposed on the pattern of soil in nature, areas of taxonomic classes rarely, if ever, coincide with mappable areas" (Soil Survey Staff Division 1993, p. 21).

It is critical that students of soil science recognize that although soil series, as a category in soil classification, have been recognized for many years, both concept and definition have changed with time. Although at one time a series was a "mappable" entity, it is now a taxonomic entity. A *mapping unit*, utilizing series names to identify the dominant soils in that unit, or soils present in a complex mapping unit, now encompasses the "mappability" criteria previously used to define series. (See Chapter 20 for discussion of mapping units.)

Differentia for series must not violate any of the differentiae of higher categories within which that series is classified. There are few specific limitations on selecting criteria for differentiating series classified within the same family. All series within the same family are referred to as "competing series." Series descriptions must fully state how each series differs from all other competing series. By not specifically limiting criteria used to differentiate among competing series, soil scientists are encouraged to be alert and respond to observations in nature where practical relationships are observed between soil properties and any soil use or natural relationship in the ecosystem. When such relationships are observed, tentative criteria are tested to determine how consistently the relationship can be observed, first in a small area and then in larger areas. It is from this constant testing that new information is injected into our understanding of soil science. If newly discovered criteria are tested and have wide applicability and value in making better interpretative statements, series differentia can become incorporated as higher category differentia.

Particle-size or texture of the surface horizon, previously considered soil type criterion, is now considered series differentia (*National Soil Taxonomy Handbook*, Issue No. 18, Unpublished 6-19-1995). Such practical features as slope of the land and presence of rocks or gravel on the soil surface are considered phases to be used in identifying mapping units. Map unit names in detailed soil surveys usually take the form of "*Cecil loam* (*3–6% slopes*)" where *Cecil* is the series name, *loam* identifies the surface texture, and *3–6% slopes* identify a slope phase. For series where the texture of the surface is included in the definition, only the series name followed by "soils" and then subsequent phases, for example, Lakeland soils, 3–6% slopes, may be used. When used in this context, the series name identifies only the taxa of the predominant soil in the mapping unit. Pedons of other series are almost always present, and the complete range of properties defined by the series is seldom fully observable within any single map unit delineation.

Perspective

Soil Taxonomy (Soil Survey Staff 1975) is the most cited single government monograph in soil science (McDonald 1994b). Since 1975 several changes have been made because the system has been used to identify soils in all parts of the world and a second edition was published (Soil Survey Staff 1999). The system has proved to be capable of incorporating new information without having to be destroyed and replaced. It is apparent that the systematic use of the connotative formative elements in the construction of the taxa names and strict observance of quantitative differentia within the keys are germane to the endurance of the system. Although a working understanding of each formative element may be a challenge to students, the formative elements provide the basis of easy and quantitative communication among soil scientists. The hierarchal structure of the system provides convenient identification of soils appropriate to any scale of map. However, it must be remembered that a taxonomic name at any categorical level in the system identifies a specific range of soil properties, and within any area of land identified by that name on a soil map, other soils will be present.

There will be changes in the system as our knowledge of soils in the world increases. It is through constant and systematic alteration of a taxonomic system, such as *Soil Taxonomy,* that new research finding can be incorporated into an easily used library of information about soils. The current system is consistent with the underlying principles and progressive nature of soil science (Haskett 1995).

Alfisols: High Base Status Soils with Finer-textured Subsoil Horizons

8

Approximately 10% of the land area of the planet is occupied by Alfisols, which because of natural fertility, location in humid and subhumid regions, and responsiveness to good management are widely used for agriculture and forestry. Alfisols occur under a wide range of environmental conditions and are found in temperate, tropical, and boreal regions of the world. The central concept of Alfisols is that of forest soils that occupy relatively stable landscape positions and thus have a subsurface zone of clay accumulation. In addition, base-rich parent materials or less-intense weathering and leaching regimes have resulted in subsoils that contain relatively abundant supplies of exchangeable calcium, magnesium, potassium, and in some cases, sodium.

Setting

Alfisols are widely distributed globally and occupy approximately 13,156,000 km^2 (USDA-NRCS database). They are found on every continent with the exception of Antarctica (Hallmark and Franzmeier 2000). Several prerequisites are met by soils of Alfisol-dominated landscapes. There is sufficient landscape stability to allow accumulation of enough layer-lattice clay (of any species) in the subsoil (often a Bt horizon) to form argillic, kandic, or natric horizons (Figures 8.1B and 8.1C). Alfisols form in parent materials to allow relatively high base (calcium, magnesium, sodium, and potassium) status, with base saturation greater than 35% in the lower part or below the argillic or kandic horizon and usually increasing with depth (Figure 8.1A). Alfisol profiles exhibit rather contrasting horizonation, which under deciduous forest typically include O, A, E, and Bt, with the possibility in various ecosystems of the presence of natric, petrocalcic, duripan, and fragipan horizons, and plinthite (Figures 8.2 and 8.3). Favorable moisture regimes provide available water to mesophytic plants more than half the year, or for three consecutive months in a warm season. Relatively little accumulation of organic matter occurs in mineral soil horizons (most organic matter is naturally cycled in the O horizons), particularly in cultivated areas (Figure 8.1B). Data illustrating these morphological, physical, and chemical properties for two Alfisols are presented in Table 8.1.

Alfisols are present on older landscapes (generally earliest Holocene or older) wherever ample supplies of primary minerals, layer-lattice clays, and available plant

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

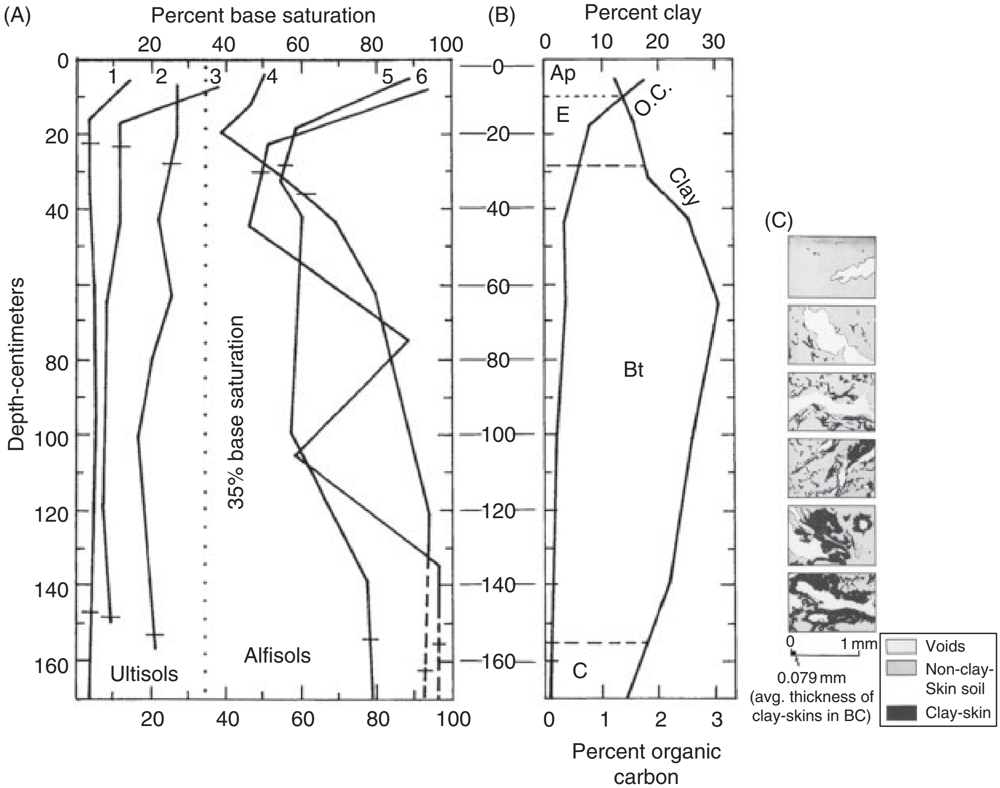

Figure 8.1. (A) Curves for base saturation, by sum of cations, for six soils: Ultisols–#1 Paleaquult (Profile #113, N.C.), #2 Fragiudult (Profile #117, Miss.), #3 Rhodudult (Profile #119, Tenn.); Alfisols–#4 Glossaqualf (Profile #45, Wis.), #5 Hapludalf (Profile #50, Ill.), #6 Haplustalf (Profile Kaduna). The *uppermost horizontal bar* on each curve marks the upper boundary of the argillic horizon; the lowermost horizontal bar marks the level 125 cm below that. *Dashed vertical lines* show unsampled portions of profiles #4 and #6. Data for first five profiles are from Soil Survey Staff 1975; for profile #6, from Harpstead 1974. (B) Curves for clay and organic carbon (O.C.), Profile #50. (C) Six tracings of representative (not average) views of thin sections from the following horizons (*from top down*; with depths of horizons in cm, and percent by volume of argillans) of a Hapludalf in Wisconsin (Buol and Hole 1961): E, 14–25 cm, 0.03%; BE, 25–48, 0.69%; Bt, 48–78, 2.69%; BC, 78–113, 3.15%; C1, 113–148, 5.27%; C2, 148–173, 2.05%. The key at lower right shows patterns to represent voids (white), soil matrix (non-clay-skin soil), and argillans (clay-skin).

nutrients are abundant in parent materials. Original vegetation included broadleaf deciduous forest, both unmixed and mixed with needle evergreen forest (North America, Europe, China); grass with and without patches of broadleaf evergreen trees (Africa, South America, California); and broadleaf evergreen forest (Africa, India, Australia, California). Areas of transition between Alfisols and Mollisols are in ecotones between forest and grassland. Transitions between areas of Alfisols and Spodosols lie in ecotones between mixed deciduous and needle evergreen forest. Under warmer temperature regimes, areas of Alfisols are found between Aridisols of

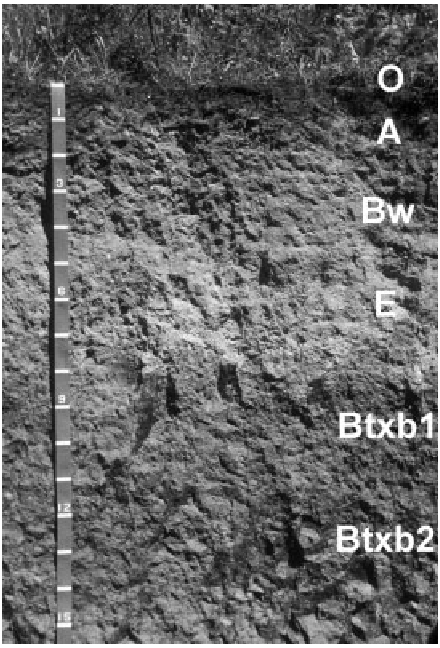

Figure 8.2. Vitrandic Fragixeralf (Santa series) from northern Idaho. Soil has formed in loess and reworked silty sediments and supports forest dominated by grand fir (*Abies grandis*). A well-developed fragipan is present a depth of ~80 cm. Scale is in decimeters. For color detail, please see color plate section.

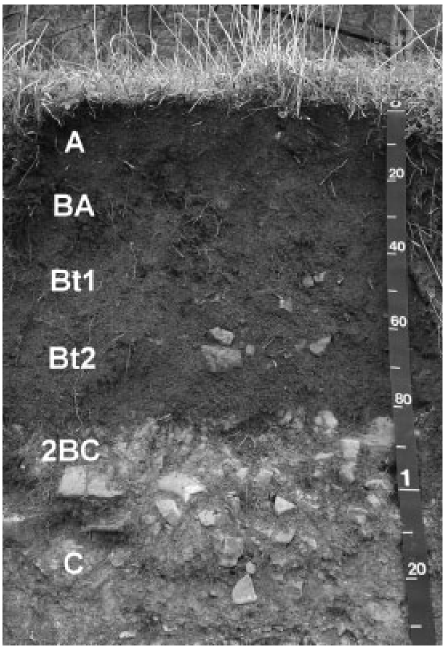

Figure 8.3. Mollic Hapludalf from Iowa County, Wisconsin. Soil has a mollic epipedon and argillic horizon; it has been leached free of $CaCO_3$. (Image courtesy of Dr. Randy Schaetzl, Michigan State University) For color detail, please see color plate section.

Table 8.1. Selected properties of Alfisols. Data for the Fayette series (Pedon no. 07N0096) and the Windthorst series (Pedon no. 96P0306) are from the USDA-NRCS National Cooperative Soil Survey Soil Characterization Database (Soil Survey Staff 2010b)

Horizon	Depth cm	Moist color	Bulk density moist g cm^{-3}	Sand	Silt	Clay	pH H_2O	C %	Exchangeable bases[a] Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	Acidity	CEC pH8	Base saturation[b] %
				%					cmol(+) kg^{1}						
Fayette series (Typic Hapludalf)—Iowa															
Ap1	0–11	10YR 3/3	1.39	9.4	72.3	18.3	6.9	1.48	13.3	1.7	–[c]	0.9	4.2	20.1	79
Ap2	11–22	10YR 3/3	1.44	9.2	70.7	20.1	6.8	1.07	12.2	2.2	–	0.5	5.1	20.0	75
Bt1	22–35	10YR 4/4	1.41	8.7	59.1	32.2	6.1	0.50	12.5	4.9	–	0.8	8.2	26.4	69
Bt2	35–54	10YR 4/4	1.32	9.6	60.4	30.0	5.8	0.36	10.9	5.4	–	0.7	9.4	26.4	64
Bt3	54–95	10YR 5/4	1.36	11.3	62.4	26.3	5.6	0.25	8.9	4.8	–	0.5	8.2	22.4	63
BC	95–142	10YR 5/4	1 .31	8.0	68.3	23.7	5.9	0.16	9.4	5.3	–	0.5	6.8	20.0	69
Windthorst series (Udic Paleustalf)—Texas															
A	0–15	7.5YR 4/4	1.42	66.7	27.4	5.9	5.7	0.84	4.1	0.8	–	0.3	2.7	7.9	66
E	15–29	7.5YR 5/4	1.54	69.5	23.8	6.7	5.4	0.31	2.6[d]	0.6	0.1	0.2	1.7	5.2	67
Bt1	29–48	2.5YR 3/4	1.34	39.0	15.7	46.3	6.6	0.53	15.0	4.7	0.2	0.9	6.3	27.1	77
Bt2	48–62	2.5YR 4/6	1.39	39.3	15.9	44.8	6.2	0.45	14.7	4.8	0.4	0.7	7.2	27.8	74
Bt3	62–83	5YR 5/4	1.52	46.8	19.1	34.1	6.4	0.27	13.5	4.4	0.4	0.5	4.8	23.6	80
Bt4	83–114	5YR 4/4	1.60	46.0	22.0	32.0	7.5	0.20	18.2[d]	5.2	0.7	0.6	2.0	26.7	93
BCk	114–140	5YR 6/6	1.41	46.4	35.3	18.3	7.9	2.46[e]	41.4[d]	3.6	0.4	0.3			100

[a]extracted with ammonium acetate.
[b]by sum of cations.
[c]none detected.
[d]may include Ca from calcium carbonate or gypsum.
[e]includes C from calcium carbonate.

the arid regions and Inceptisols, Ultisols, and Oxisols of the more humid areas (Soil Survey Staff 1999). Transitions between Alfisols and Aridisols may also occur in ecotones between cool, temperate shrub/grassland and drier steppe vegetation in the western United States (Blank and Fosberg 1991).

In cooler regions affected by glaciation, Alfisols are generally found on late Pleistocene surfaces (Hallmark and Franzmeier 2000). Under warmer, subhumid and semi-arid conditions, Alfisols occupy older, stable landscapes that have undergone long-term weathering under fairly constant climatic conditions or are polygenetic, and have experienced climatic changes during the Pleistocene. Under humid, temperate climates, Alfisols may occupy most of the landscape except for very steep slopes, alluvial floodplains, and very poorly drained depressions. They also are common on the borders of depressions where slight concentrations of water have favored migration of sodium and clay to form natric horizons. Where Mollic Albaqualfs are associated with Mollisols in Illinois, their presence has been attributed to lower organic matter production caused by nutrient deficiencies associated with a detrimental moisture regime (Smeck and Runge 1971). Higher elevations, limited rainfall, and higher concentration of bases in parent rock favor Alfisol formation in the tropics (Guerrero 1963). Alfisols can also occur in highly weathered tropical landscapes on dissected side slopes where less-weathered parent materials are exposed (Lepsch et al. 1977b). These Alfisols exist in areas that have been geomorphically 'rejuvenated' on an otherwise highly weathered landscape (Schaetzl and Anderson 2005).

Pedogenic Processes

The dominant processes in most Alfisols are those that produce subsurface horizons relatively enriched in layer silicate clays and those that cycle nutrients in upper horizons, mostly by littering and organic matter decomposition. Although the genetic concept of argillic and natric horizon formation emphasizes to some degree the process of clay translocation (lessivage), the formation of these horizons and the kandic horizon involves other processes as well. These processes include the selective loss of clay from A and E horizons by dispersion and lateral transport, dissolution of clays in A and E horizons and leaching of dissolved constituents, neosynthesis of clays in the B horizon from dissolved constituents, clay production in the B horizon by weathering of primary minerals, and residual concentration of clay by selective dissolution of more soluble minerals (e.g., loss of calcium carbonate in calcareous alluvium, loess, or till).

Translocation of clay from the A and E horizons into the B horizon, in an aqueous suspension (lessivage), with or without aid of complexing organic compounds, and possibly by migration of Si and Al and their later precipitation in the B horizon (neoformation), probably does play a major role in the formation of some argillic horizons and natric horizons (Thorp et al. 1957, 1959). Fine clays move more readily than do coarse clays, so the fine clay-to-total clay ratios are typically higher in the B horizon (0.6–0.8) than in the A and E horizons (0.3–0.6 [Isbell 1980]). This distinguishes argillans from rapidly formed alluvial coatings, called gleyans by Brammer (1971), in certain Fluvents. Scarcity of argillans in the upper B horizon (Figure 8.1C) is considered to be a result of their destruction at a faster rate than that of their formation. Shrink-swell cycles (by both freeze-thaw and wet-dry processes) (Nettleton et al. 1969), soil creep, and biologic mixing (Vepraskas and Wilding 1983) are more intense in the upper horizons where argillans are few, and where they are largely embedded inside peds (Figure 8.1C). Interfingering of albic material into the argillic or natric horizon, and in extreme cases the formation of a glossic horizon, are examples of the destruction of the upper part of argillic or natric horizons (Ranney and Beatty 1969), and disruption of not only argillans, but of the entire clay-enriched soil fabric. Grainy ped coatings observed in the upper part of the argillic horizon of degrading Udolls (Arnold and Riecken 1964) may represent a first stage in this process.

Deposition of clay, often with sesquioxides and organic matter, in the argillic horizon may be brought about by (1) depletion of percolating waters via sorption by peds, (2) swelling shut of voids and consequent slowing of percolating water, (3) sieve action by clogging of fine pores, and (4) flocculation of the negatively charged clay by positively charged iron oxides in the Bt horizon or by calcium in the higher-base-saturation lower solum. The clay may have originated in eluvial (A or E) horizons, may have formed by crystallization of soluble constituents produced by weathering elsewhere in the pedon, or may have been produced locally by weathering in the B horizon. Conditions for significant formation of argillans in Alfisols may be relatively rare, with intensely rainy periods following prolonged droughts being especially conducive. The laminar structure of some argillans, visible in thin sections, reflects episodic deposition of mineral and organic fines.

Considering a ped of the argillic horizon as a pedologic "unit cell," we may expect that the microclimate is different in the interior from that on the ped surface (Heil and Buntley 1965; Buntley and Westin 1965). An argillan acts as a barrier to penetration of water, nutrients, and growing roots as well as to the movement of soil fauna into and out of the ped (Khalifa and Buol 1969).

The evolution of pedons of Paleustalfs in southeastern Australia (Walker and Chittleborough 1986) illustrates the formation of A and E horizons that are largely composed of silt and sand in the inorganic fractions and B horizons that have become more bimodal in texture as clay has been added to the original material. With progressive blocking of voids by illuvial clay, seasonal waterlogging has become more frequent in the upper solum. Weathering, a process that began at or before time zero, has become predominant over translocation, although this interpretation is complicated by evidence of eolian addition to soils of the region (Walker et al. 1988) and by possible stratification of the initial geologic material.

Munk's (1993) study of a sequence of soils formed from 2000 to 200,000 year-old alluvium in California's Central Valley also suggests that a number of mechanisms are responsible for clay accumulation in the Alfisols (Figure 8.4). Haploxeralfs form in latest-Pleistocene to early Holocene deposits by illuviation and in situ mineral weathering. Plugging of voids in deeper horizons causes clay to accumulate above the

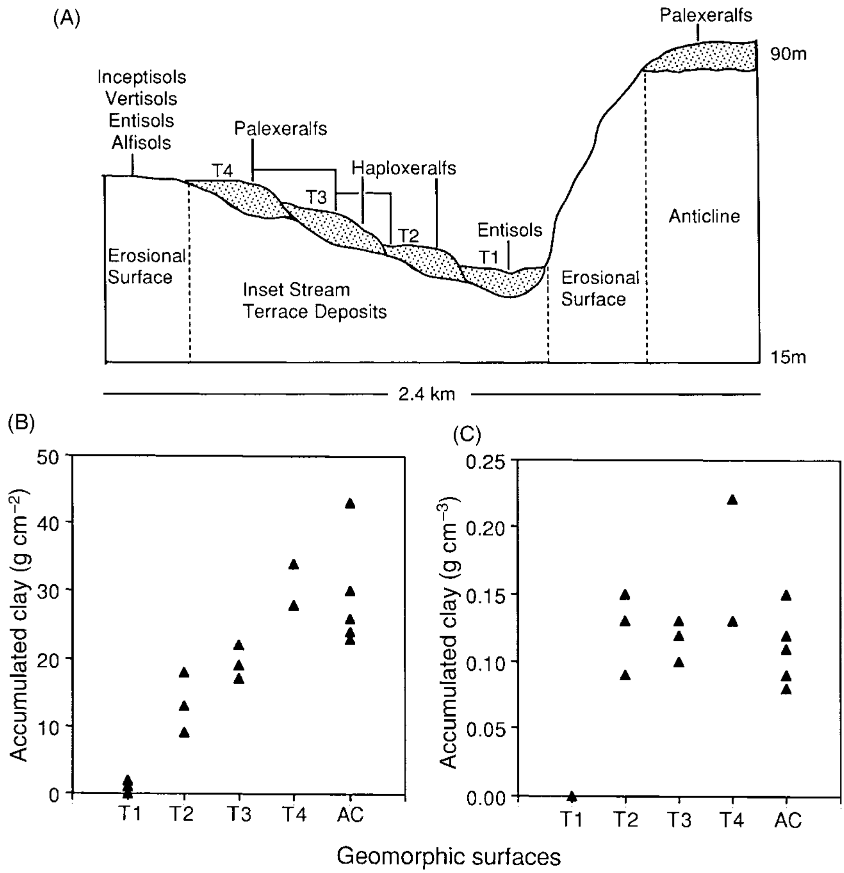

Figure 8.4. Geomorphic relationships and clay accumulation in Xeralfs and associated soils in central California. (A) Idealized cross section of geomorphic relationships among stream terrace deposits (T1, T2, T3, and T4) and the dissected anticline. (B) Total accumulated clay in soils on terraces and the anticline cap (AC). (C) Accumulated clay on a depth-weighted basis (accumulated clay shown in B, divided by solum thickness). Triangles show the range of values for soils on each surface. (After Munk 1993)

zones of slower permeability (in the initial stages, the argillic horizons grow from the bottom up). The slower permeability also favors the in situ weathering of primary minerals directly to clays and the retention of weathering products that favors clay neosynthesis. Palexeralfs, with abrupt textural changes at the A-B horizon contact, form on earlier-Pleistocene deposits when clay accumulation and slow permeability are sufficient to cause perching of a seasonal water table in the winter (epiaquic conditions), thereby leading to destruction of clay by ferrolysis. The increased clay contents of soils formed on progressively older deposits (Figure 8.4B) are initially a result of illuviation. Thickening of the solum by clay formation in the B horizon is mostly responsible for subsequent clay accumulation (Figure 8.4C).

In the north central United States, weathering of highly calcareous glacial till has yielded exceptionally clayey argillic horizons by concentrating the clay as an insoluble residue, once the carbonates have been leached out (Franzmeier et al. 1985). Weathering has also reduced the size of sand and silt particles in Alfisols, particularly in the upper solum. It has converted some of those skeleton grains to clay papules that can be distinguished from chips of shattered argillans and has reduced clay particle size. Under seasonally alternating reducing and oxidizing conditions, ferruginous nodules >2 mm in diameter have formed, in the A and B horizons first, and later in the C horizon. A field and laboratory study of a family and catena of Alfisols in the north central United States (Brown and Thorp 1942) also showed that iron oxide concretions form and accumulate in horizons affected by a seasonally perched water table above dense B horizons (Figure 8.5).

Leaching of carbonates from the zone of the developing solum and braunification appear to be prerequisites to the development of many Alfisols. The presence of exchangeable calcium (from calcium carbonate) flocculates clay particles, creating particles that are too large to be physically transported in suspension. Removal of the calcium leaves the solum in a condition favorable to relatively free movement of plasma under the influence of percolating water. Soil colloids apparently move in the acidic soil environment of some Hapludalf Bt horizons, where soil reaction may be as acid as pH 4.5. In Alfisols with natric horizons, wherein subsoil pH may be as high as 10.5, the sodium ion is important in the dispersion and mobilization of clay. In soils where the total solute concentration is low, clay dispersion can occur when exchangeable sodium percentage (ESP = 100[Na/CEC]) is as low as about 5 (Shainberg and Letey 1984). Movement of colloids by cheluviation does occur, as suggested by the presence of striking albic and argillic horizons in Eutrudalfs, which may be nearly neutral to the surface. Braunification releases another flocculant, iron,

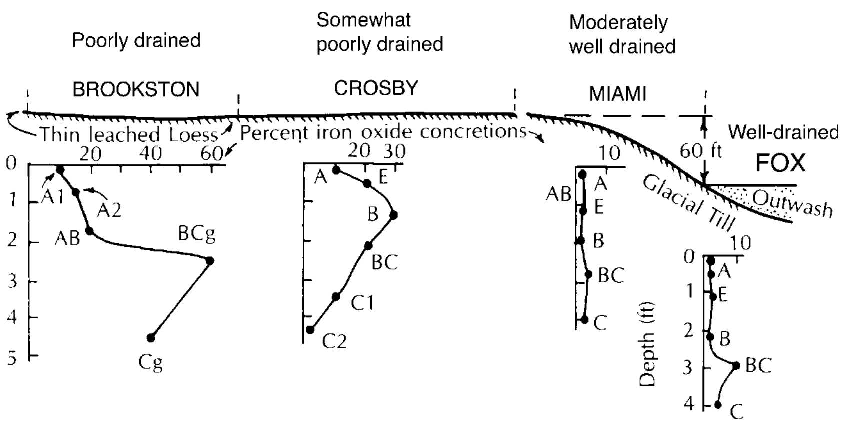

Figure 8.5. Distribution of iron concretions in sand and gravel fractions of soils making up an Alfisol-dominated catena. The more well drained soils of the catena, Miami and Fox, are both Hapludalfs and contain relatively few concretions. The most poorly drained soils, Brookston, are Argiaquolls. (After Brown and Thorp 1942)

which fosters deposition of clay in the main Bt horizon, whereas, carbonates of the Bk or C horizon flocculate colloids near the base of the solum.

The process of lessivage in combination with other soil-forming processes can result in formation of bisequal Alfisols. An example of this is seen in Udalfs of northern Michigan (Schaetzl 1996). Translocation of clay from E to Bt horizons gives rise to relatively coarser-textured E horizons; the coarser textures, in turn, help promote podzolization, and a Spodosol-like E-Bs horizon sequence forms within the E horizon (Schaetzl 1996). The net result is bisequal profile consisting of an A-E-Bs-E′-Bt-C horizon sequence. These bisequal soils represent the transition between Alfisols and Spodosols (Schaetzl 1996). If podzolization is sufficiently well expressed, soils are classified as Spodosols rather than Alfisols; the Trenary series (Alfic Haplorthods) is one such example (Soil Survey Division 2010).

The Oi, Oe, and Oa are dynamic horizons that exhibit an annual cycle of winter maximum accumulation and late summer minimum in some Hapludalfs under forest cover (Nielsen and Hole 1964) (Figure 8.6). Processes such as tree tip are probably unimportant in the genesis of Alfisols. Tree-tip mounds are scarce in originally forested Alfisols as compared with Spodosols, probably because rooting is commonly deeper in the Alfisols.

Formation of an ochric epipedon has been almost universal in these soils, whether currently under prairie or forest, although umbric and mollic epipedons do occur in Alfisols (Figure 8.3). Roots of plants are primarily responsible for melanization in grassland soils through the in situ addition of organic residues. Where this process is not sufficient to form Mollisols with mollic epipedons, Alfisols have formed under prairie cover. This generally seems to be the case in thermic and hyperthermic soil temperature regimes where organic matter decomposition is rapid. In forested ecosystems, the trees deliver the bulk of their annual production of organic matter aboveground, including the litter that is not as deeply incorporated as in grass-covered soils. Thus,

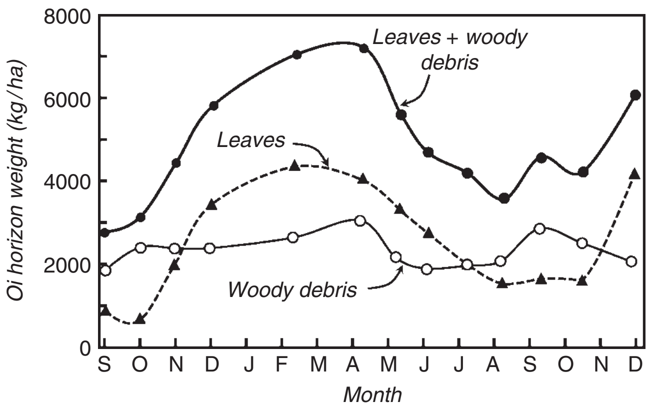

Figure 8.6. Seasonal changes in dry weight of organic horizons of a Hapludalf under forest. (After Nielsen and Hole 1964).

the impression that the dark Mollisols of grasslands contain more organic matter than the Alfisols of nearby forestlands, may not reflect the large amount of organic matter in O horizons and layers above the ochric epipedon of the Alfisols. In forestlands, additional work in the genesis of the A horizon is done by earthworms (Nielsen and Hole 1964) and other fauna that mix leaf litter and humus with mineral soil to shallow depths (2–10 cm). Extracellular substances (e.g., polysaccharides) excreted by microbes involved in the decomposition of organic materials in the O and A horizons are important in the formation of granular structure in surface mineral horizons (Chenu 1995). Biocycling of nutrients from B horizons to O and A horizons is an important process in most forested Alfisols. This process is indicated by the profile distribution of base saturation (Figure 8.1A) and soil reaction: nearly neutral (pH 6.5–7.0) surface soil (A horizon) over quite acidic (pH 4.8–5.8) subsoil. Residues of calcareous concretions from tissues of earthworms have been observed in A horizons of forested Alfisols (Wiecik and Messenger 1972). The changing microtopography of the forest floor can be related to seasonal activities of animals, including moles and rodents.

Fragipans are common in many Alfisols of humid, temperate regions (Hallmark and Franzmeier 1999) and are also locally extensive under xeric moisture regimes (McDaniel and Falen 1994) (Figure 8.2). Given the range in environments in which they are found, there are probably a variety of genetic pathways via which fragipans can form. Possible mechanisms of formation include physical ripening, collapse of loess upon wetting and drying, compaction by the weight of glaciers, permafrost processes, and chemical cementation (Smeck and Ciolkosz 1989; Soil Survey Staff 1999; Hallmark and Franzmeier 1999). Regardless of genesis, fragipans are characterized by low hydraulic conductivity and high strength when dry. These features play an important role in limiting root development and vertical percolation of water. By restricting vertical water movement, fragipans cause formation of perched water tables or *episaturation*. This can result in saturation-excess runoff (Needelman et al. 2004; Gburek et al. 2006) as well as significant subsurface flow and redistribution of water and solutes within watersheds (Hammermeister et al. 1982; Reuter et al. 1998) (Figure 8.7). As much as 90% of the incident late winter/early spring precipitation has been shown to leave Fragixeralf hillslopes as subsurface lateral flow (McDaniel et al. 2008).

Albic horizons overlie the argillic, fragipan, or natric horizons of many Alfisols. The effects of leaching and eluviation of clay are most obvious in the albic horizon where melanization has been minimal. Platy structure is common in albic horizons of soils of the cooler regions and may result from expansion and contraction normal to the soil surface, in freeze-thaw and wetting-drying cycles, and from parallel orientation of planar silt particles. A seasonally perched water table may be involved. Vesicular structure may result from degassing of water on freezing. Physical breakdown of fine sand and coarse silt may be most intense in the albic horizon. Ferrolysis and subsurface lateral flow is apparently responsible for decreased clay content also found in many of these horizons (McDaniel et al. 2001). Some albic horizons are relatively high in colorless or light-colored organic matter (Wilde 1950). Resistant

Figure 8.7. Effect of a fragipan on near-surface hydrology in a cultivated Fragixeralf landscape in northern Idaho. A seasonal perched water table above the fragipan appears as the dark layer in the road cut. Much of the incident precipitation is shed from these hillslopes as lateral throughflow.

minerals such as quartz are more concentrated in the A horizons, and the SiO_2:R_2O_3 ratio is higher than in the Bt (Allan and Hole 1968).

Uses of Alfisols

Alfisols are used for cultivated crops, winter (hardy) hayland, pasture, range, and forest. The relatively high base saturation of most pedons and the presence of large reserves of plant nutrients in the more highly base-saturated C horizon indicate the native fertility of these soils. They support the eastern portion of the Corn Belt (in association with Cumulic Hapludolls and Aquolls [Figure 8.8]) and the western part of the wheat-producing areas of North America. Arable soils of Mediterranean climatic zones are to a considerable extent Alfisols. Wine grapes are grown extensively on Xeralfs throughout the world.

Accelerated erosion presents a serious hazard to future productivity of Alfisols. The argillic, natric, or kandic horizon, being relatively high in clay content, is not a very desirable medium for seed germination and plant development when exposed at the soil surface. Infiltration may be decreased on an area from which ochric and albic horizons have been lost by erosion. Accelerated erosion not only fosters a droughty condition at the eroded site but also increases flood hazard in lower portions of the adjacent landscape.

The abundance of relatively unweathered clay minerals provides many temperate zone Alfisols with a high cation exchange capacity (CEC), and aluminum toxicity is not usually a problem. In the tropics, many and perhaps most Alfisols have low CEC kandic

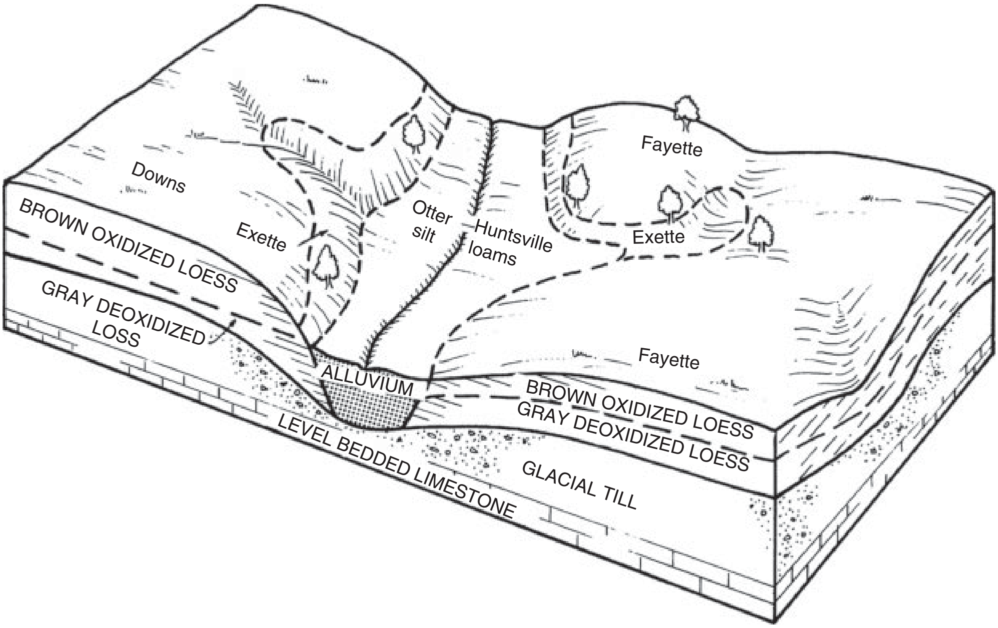

Figure 8.8. Soilscape pattern in Fayette County, Iowa. Alfisols (Fayette and Downs) occupy the stable upland positions. Fayette soils are Typic Hapludalfs and formed under oak-hickory forest. Downs soils are Mollic Hapludalfs and formed under prairie grasses and open oak-hickory woodlands. Exette soils (Dystric Eutrudepts) are found on the steeper sideslopes, and Mollisols (Huntsville and Otter) have formed in valley alluvium. Huntsville soils are Cumulic Hapludolls, and Otter soils are Cumulic Endoaquolls. All of the soils are in fine-silty, mixed, superactive, mesic families. (After Kuehl and Highland 1978)

horizons, and aluminum toxicity in the A and upper Bt horizons is a frequent problem. Potassium and ammonium fixation is an important consideration in Alfisols that have notable contents of vermiculite and slightly weathered mica in the clay fraction.

Classification of Alfisols

Alfisols have an argillic, kandic, or natric horizon that is not under a spodic or oxic horizon. Usually, an ochric epipedon is present. The base saturation by sum of cations is at least 35% at a depth of 125 cm below the upper boundary of the argillic, natric, or kandic horizon (Figure 8.1A) or 180 cm below the mineral soil surface. If a fragipan is present, clay films (argillans) at least 1-mm thick must be present in some part of the fragipan, and base saturation must be at least 35% at a depth 75 cm below the upper boundary of the fragipan, or within 200 cm of the mineral soil surface. Where a lithic or paralithic contact is present within one of the aforementioned depths, a base saturation of at least 35% is required at the contact. Any soil temperature regime is allowed, except gelic. Alfisols do not have an aridic soil moisture regime.

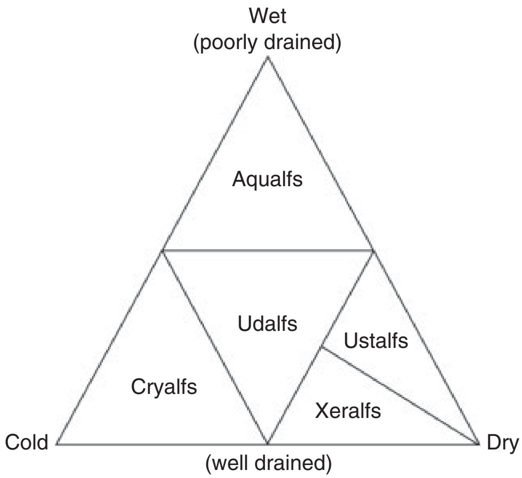

Figure 8.9. Diagram showing some relationships among suborders of Alfisols.

Alfisols are divided into five suborders (Figure 8.9), which are briefly described below.

Aqualfs have aquic conditions for some time in most years within 50 cm of the mineral soil surface and redoximorphic features in the upper 12.5 cm of the argillic, natric, or kandic horizon. These soils are found in a wide range of climatic conditions and exhibit a variety of soil horizons and features. They frequently occupy depressional areas or low-gradient landscapes subject to seasonal inundation or high water tables.

Cryalfs are the cold Alfisols that have either a cryic or isofrigid temperature regime. Most Cryalfs have a udic moisture regime, and are generally found above 49° N. latitude in North America, Europe, and Asia (Soil Survey Staff 1999). They may also be found at higher elevations at lower latitudes. Cryalfs occupy about 1.9% of the global land area (as shown in Table 20.3) and include many soils formerly classified as Boralfs in earlier versions of Soil Taxonomy.

Udalfs are not as wet as Aqualfs, nor as cold as Cryalfs. These soils are extensive in the midwestern and eastern United States, in northern Europe, and in some coastal regions of Africa and Australia (Figure 8.3). Selected data for a Udalf from Iowa are provided in Table 8.1.

Ustalfs have ustic soil moisture regimes and are the most-extensive suborder of Alfisols, occupying approximately 4.6% of the world's ice-free area. (See Table 20.3.) Many Ustalfs have a calcium carbonate accumulation at the base of the solum. Ustalfs occupy large areas of the tropics in Africa, India, South America, Australia, and southeastern Asia. In the United States, Ustalfs occur mostly in Texas, Oklahoma, and New Mexico. Selected data for an Ustalf from Texas are provided in Table 8.1.

Table 8.2. Suborders and Great Groups in the Alfisol Order

Suborder	Great Group
Aqualfs	Cryaqualfs have a cryic soil temperature regime.
	Plinthaqualfs have plinthite between 30 and 150 cm.
	Duraqualfs have a duripan.
	Natraqualfs have a natric horizon.
	Fragiaqualfs have upper surface of fragipan within 100 cm of the surface.
	Kandiaqualfs have a kandic horizon.
	Vermaqualfs have 25 cm thick layer within 100 cm of surface with 50% or more (by volume) recognizable bioturbations.
	Albaqualfs have abrupt textural change between ochric epipedon or albic horizon and argillic horizon with moderately low or low hydraulic conductivity.
	Glossaqualfs have a glossic horizon.
	Epiaqualfs have episaturation.
	Endoaqualfs have endosaturation (other Aqualfs).
Cryalfs	Palecryalfs have loamy fine sand or finer argillic, kandic, or natric horizon with upper boundary at or below 60 cm and a glossic horizon or interfingering of albic materials.
	Glossocryalfs have a glossic horizon.
	Haplocryalfs—other Cryalfs.
Ustalfs	Durustalfs have upper boundary of duripan within 100 cm.
	Plinthustalfs have some horizon with one-half or more plinthite within 150 cm.
	Natrustalfs have a natric horizon.
	Kandiustalfs have a kandic horizon within which clay content does not decrease more than 20% of its maximum within 150 cm.
	Kanhaplustalfs have a kandic horizon.
	Paleustalfs have the upper boundary of a petrocalcic horizon within 150 cm, or an argillic horizon within which clay content does not decrease more than 20% of its maximum within 150 cm.
	Rhodustalfs have 2.5YR or redder colors with moist values of 3 or less that increase no more than one value unit when dry in 50% or more of all horizons to 100 cm.
	Haplustalfs—other Ustalfs.
Xeralfs	Durixeralfs have upper boundary of a duripan within 100 cm.
	Natrixeralfs have a natric horizon.
	Fragixeralfs have upper boundary of a fragipan within 100 cm.
	Plinthoxeralfs have some horizon with one-half or more plinthite within 150 cm.
	Rhodoxeralfs have 2.5YR or redder colors with moist values of 3 or less that increase no more than one value unit when dry in 50% or more of all horizons to 100 cm.
	Palexeralfs have the upper boundary of a petrocalcic horizon within 150 cm, or an argillic or kandic horizon within which clay content does not decrease more than 20% of its maximum within 150 cm.
	Haploxeralfs—other Xeralfs.
Udalfs	Natrudalfs have a natric horizon.
	Ferrudalfs have a glossic horizon and discrete iron-enriched nodules, 2.5 to 30 cm in diameter in argillic or kandic horizon.
	Fraglossudalfs have a glossic horizon and the upper boundary of a fragipan within 100 cm.
	Fragiudalfs have upper boundary of a fragipan within 100 cm.

Table 8.2. *Concluded.*

Suborder	Great Group
	Kandiudalfs have a kandic horizon within which clay content does not decrease more than 20% of its maximum within 150 cm.
	Kanhapludalfs have a kandic horizon.
	Paleudalfs have an argillic horizon with hues of 7.5YR or redder and within which clay content does not decrease more than 20% of its maximum within 150 cm.
	Rhodudalfs have 2.5YR or redder colors with moist values of 3 or less that increase no more than one value unit when dry in 50%or more of all horizons to 100 cm.
	Glossudalfs have a glossic horizon.
	Hapludalfs—other Udalfs.

Xeralfs have xeric soil moisture regimes. Like the Ustalfs, many Xeralfs have carbonate accumulations in the lower solum. The Xeralfs are found in southern Europe in areas around the Mediterranean Sea, South Africa, western United States, Chile, and southern Australia (Soil Survey Staff 1999).

The suborders are divided into 39 great groups (Table 8.2) on the basis of a number of soil properties, including interfingering of albic material into the argillic horizon, abrupt textural change into the argillic horizon, thickness of the argillic horizon, reddish colors in the argillic horizon, presence of low activity clays, presence or absence of a fragipan (Figure 8.2), duripan, agric, glossic, natric, or petrocalcic horizon or umbric epipedon, presence or absence of iron nodules or plinthite, percent base saturation in the argillic horizon, and the nature of saturated conditions (episaturation versus endosaturation).

Subgroups in the Alfisol order reflect the relationship of these soils with several other orders that border the modal Alfisol ecological niche. Intergrades to the Mollisol, Ultisol, and Vertisol orders are the most common. Properties transitional to Aridisols, Entisols, Inceptisols, and Andisols are also recognized, as are a number of extragrade properties (e.g., Lithic, Arenic, Aeric).

Drainage differences commonly noted in a catena of Alfisols are reflected in the suborder and subgroup classification. (See Figure 8.5.) For example, in a typical loess-over-glacial till landscape of the north central United States, well-drained soils are Typic Hapludalfs (Fox series); moderately well drained soils are Oxyaquic Hapludalfs (Miami series). The somewhat poorly drained soils are Aeric Epiaqualfs (Crosby series), and the poorly drained soils are Typic Argiaquolls (Brookston series).

The majority of Alfisols with higher-activity clays are equivalent to the Luvisol and Albeluvisol reference soil groups in the *World Reference Base for Soil Resources* (IUSS Working Group WRB 2006). Alfisols with low-activity clays would be included in the Nitisol and Lixisol soil reference groups in WRB; those with plinthite would be included with Plinthosols.

Perspective

Alfisols may be thought of as morphologically developed, naturally forested soils, on relatively fertile parent materials. Principles of pedogenesis and land management have been notably advanced by studies of interactions among the distinct horizons (O, A, E, B), among members of topo-drainage-sequences, among soils of developmental weathering sequences, and between these soils and the plants that they support. It is little wonder that Alfisols support productive agriculture and silviculture over wide areas around the world.

In summary, the following processes contribute to the formation of Alfisols:

1. Cycling of nutrients from the subsoil through plants to O and A horizons mainly via littering and biological decomposition.
2. Residual concentration of clay in the B horizon by the removal of soluble minerals (especially calcium carbonate) from the initial material of that horizon.
3. Eluviation, from the A and E horizons, of clay in the initial material, of clay formed by mineral weathering, and of clay progressively added in eolian material, and subsequent illuvial accumulation of the clay in the B horizon.
4. Formation of clay in the B horizon by in situ weathering of feldspars, micas, and ferromagnesian minerals, or by neosynthesis from illuvial weathering products.
5. Selective loss of clay from A and E horizons by dispersion and lateral transport or by ferrolysis.
6. Differential loss by leaching of materials, so that the A and E horizons are more depleted, particularly in clay, than is the B horizon, but both are depleted.

Andisols: Soils with Andic Soil Properties

9

Andisols are soils with properties dominated by short-range-order compounds such as aluminosilicates, ferrihydrite, and organometallic complexes. The vast majority of Andisols formed from volcanic tephra (ash, pumice, cinders) and related volcanic parent materials, although a few have formed from nonvolcanic materials. Formally adopted as the 11th soil order in 1990 (Soil Survey Staff, 1990), Andisols represent many soils known by other names, including Ando soils, Kurobokudo, Andosols, and Volcanic Ash Soils. Although the definition does not precisely match the definitions of these other groups of soils, Andisols are closely related to Andosols, one of the 32 reference soil groups in the World Reference Base for Soil Resources (IUSS Working Group WRB 2006). In the first edition of Soil Taxonomy, Andisols were mostly identified as Andepts and Andaquepts (Inceptisols), a placement that recognized the properties resulting from minimal crystallization and redistribution of weathering products.

Setting

The primary control of Andisol characteristics is volcanic parent material, especially ash. Volcanic ash refers to the <2-mm-diameter material that is blown out of a volcano during an eruption. Thus, the geographic distribution of active volcanoes and those that have been active during the Holocene and, in some cases, Late Pleistocene is the major determinant of Andisol distribution. Soils formed on volcanic deposits older than these generally have mineral suites dominated by crystalline aluminosilicates. Parent materials of Andisols encompass a wide range of chemical composition and mineralogy, and include lava flows, tephras, pyroclastic flows, lahar deposits, volcanic alluvium, and volcanic loess, but generally exclude ultramafic volcanic material (Neall 1985). It is estimated that Andisols only occupy approximately 0.7 of the earth's ice-free land surface, or just under 963,000 km^2, making it the least extensive of any of the soil orders (Soil Survey Staff 1999). Major areas occur from Iceland in the Northern Hemisphere to southern Chile in the southern hemisphere (Leamy et al. 1980). Extensive "belts" are found around the "ring of fire" in the Pacific including the Andes of South America, the Central America cordillera, Alaska, Japan, Philippines, Indonesia, New Zealand, and the Pacific Northwest of the United States

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

(McDaniel et al. 2010). Small areas of The Rift Valley of Africa, as well as Italy, France, Spain, and Romania also have important areas of Andisols.

All soil moisture regimes and all soil temperature regimes are found in Andisols. However, limited water for hydrolytic reactions in extremely arid regions prevents the weathering of volcanic ash, and hence the formation of Andisols. Vegetation is also very diverse, ranging from desert shrubs in arid regions, to dense coniferous forests in humid regions, to tundra in cold regions of higher latitudes and elevations. The association of Andisols with volcanic activity also dictates that they are frequently on steep mountain slopes at higher elevations. However, forming from airborne material, they can be found on any terrain, including floodplains where the volcanic material is water deposited.

Pedogenic Processes

Many of our usual concepts of soil formation have to be modified somewhat when considering the genesis of Andisols. The most noticeable difference is that the youngest and sometimes least-weathered part of the soil is the surface layer or horizon. Many Andisols are formed in unconsolidated volcanic ejecta originating from sequential eruptions of one or more volcanoes. The result is a distinctly layered soil profile with buried soil horizons (Figure 9.1). In quiet periods between eruptions, weathering and other soil-forming processes proceed in a top-down manner, giving rise to a sequence of soil horizons. A subsequent eruption can then cover the land surface with fresh tephra. In cases where tephra accumulates rapidly, existing soil horizons may be deeply buried and effectively isolated from further pedogenesis as soil formation resumes at the new land surface. This scenario is known as retardant upbuilding (Schaetzl and Anderson 2005) and gives rise to soils such as the one shown in Figure 9.1. Alternatively, if subsequent eruptions produce relatively thin additions of new tephra and accumulation rates are low, developmental upbuilding continues (Schaetzl and Anderson 2005). As new tephra is added to the soil profile, it is exposed to pedogenesis before burial. This results in the intermixed tephra deposits having a soil fabric inherited from when the tephra was part of the A or B horizon (McDaniel et al. 2010). In short, understanding Andisol genesis in many instances requires a stratigraphic approach in combination with an appreciation of buried soil horizons and polygenesis.

The nature of ash falls varies greatly both in time and space. Volcanic events can be of short duration or persist as a continuous eruption over several years. The size of particles deposited at a site may one day be coarse and the next fine, depending upon the direction and speed of the wind. Some eruptions completely bury both soil and vegetation, whereas many and perhaps most ash falls only partially bury the vegetation, which continues to grow as the ash accumulates. Materials of contrasting composition are common from the same volcano, even during a single eruption.

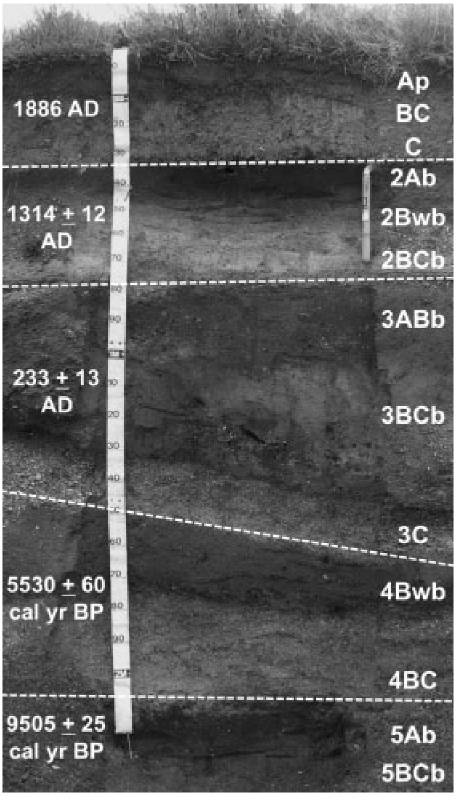

Figure 9.1. Typic Udivitrand from central North Island of New Zealand. Exposure shows sequence of buried soil horizons formed in multiple tephra deposits. Stratigraphy, tephra unit ages (left side of figure), and horizon designations are from Lowe (2008). Scale divisions = 10 cm.

Volcanic ash is mineralogically different from most other parent materials. Rapid cooling of the molten materials upon ejection prevents crystallization of minerals with long-range atomic order, and the resulting product is known as volcanic glass or vitric material (Figure 9.2). Although the bulk chemical composition may be similar to that of other rocks with well-defined mineral constituents (e.g., rhyolitic, andesitic, basaltic), discrete minerals do not exist. As a result, volcanic glass is more weatherable than crystalline materials with the same bulk composition. Volcanic eruptions that produce significant amounts of ash are usually explosive due to the high silica content of the magma and a large component of gases (water, carbon dioxide, hydrogen sulfide). The explosive eruptions consist not only of glass formed by the cooling of magma, but also of fragments of rocks, often hydrothermally altered, from the throat of the volcano. Many of these lithic fragments may be coated by glass. The resulting ash fall, often called tephra deposits or pyroclastic material (approximate synonyms), therefore may be a mixture of crystalline minerals and volcanic glass.

The glassy materials that dominate the coarser fractions of many Andisols weather relatively rapidly to form colloidal materials that possess short-range order. Once

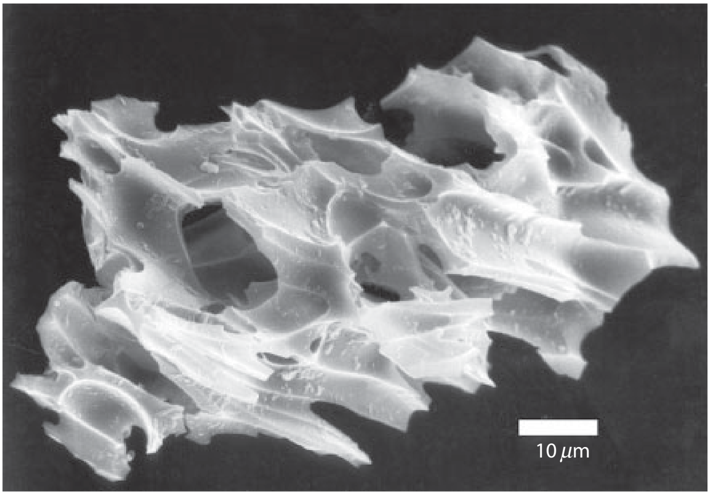

Figure 9.2. Scanning electron micrograph of a glass shard from the eruption of Mount Mazama (now Crater Lake, OR) approximately 7,700 years ago. The vesicles seen in the glass indicate a highly explosive eruption of viscous magma. Today, this glass blankets most of the forested regions of the Pacific Northwest. (Image courtesy of University of Idaho.)

thought to be amorphous, these materials comprise very small but structured nanominerals (Hochella 2008). The predominant weathering products include allophane, imogolite, ferrihydrite, and metal-humus complexes. These minerals differ markedly from those of most other mineral soils and confer the unique properties associated with Andisols. Even a fundamental soil property such as texture takes on a new meaning, as many of the weathering products, especially in moister environments, tend to exist as gels rather than discrete clay-size particles. This, coupled with high variable charge, makes complete dispersion of these materials difficult. Not surprising is the fact that laboratory particle-size analyses generally indicate less clay-size material than do field estimates (Ping et al. 1989).

The suite of weathering products found in Andisols depends to a large extent on the leaching regime, the acidity of the weathering environment, the supply of organic acids, and the presence of 2:1 layer silicates. Shoji et al. (1993) and Dahlgren et al. (2004) provide extensive discussions of the weathering environments and processes. Brief descriptions of two of the most common weathering scenarios are presented here.

As glass and other minerals weather via dissolution and hydrolysis by carbonic acid, aluminum and silicon are released. Under conditions where soil pH ~>5 or when organic matter production is relatively low, aluminum and silicon polymerize and precipitate to form the short-range-order aluminosilicates allophane and imogolite in surface horizons. Iron may also precipitate as ferrihydrite, a short-range-order oxyhydroxide (Bigham et al. 2002). These weathering products are characteristic of *allophanic* Andisols (Figure 9.3).

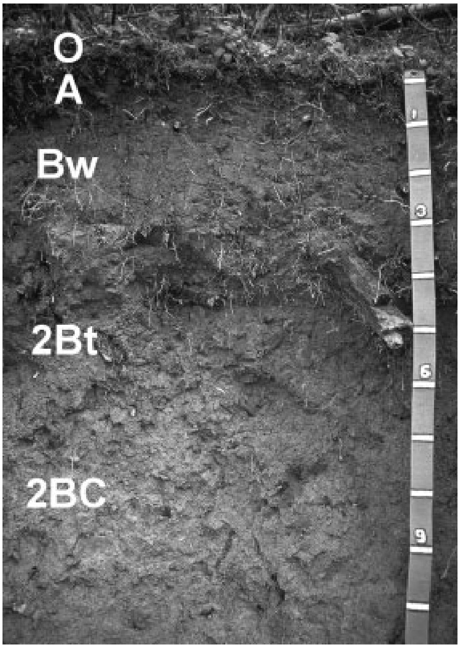

Figure 9.3. Alfic Udivitrand from northern Idaho. Soil has formed in 7,700-year-old ash from the cataclysmic eruption of Mount Mazama (now Crater Lake, Oregon). Volcanic ash mantle extends to a depth of ~40 cm. Scale divisions are decimeters. For color detail, please see color plate section.

Although once considered amorphous, allophane consists of hollow spheres with diameters of 3.5–5 nm (Parfitt 2009) (Figure 9.4A). Imogolite has a somewhat similar chemical composition but appears as long thread-like tubes with inner and outer diameters of 1 and 2 nm, respectively (Churchman 2000) (Figure 9.4B). In some cases, poorly crystalline smectites appear to be the initial weathering products, which subsequently are desilicated to form allophane and imogolite (Southard and Southard 1989). Over time, dehydration and structural rearrangement of allophane and imogolite may lead to halloysite formation, especially if desilication is not too severe. Halloysite formation is favored under conditions of lower rainfall, Si-rich parent materials, and restricted drainage (Lowe 1986; Churchman 2000). Progressive desilication may ultimately produce gibbsite in some Andisols, although the horizons that contain gibbsite may not have andic properties. Seasonal dryness, as in ustic and xeric soil moisture regimes, seems to speed the formation of the crystalline clays, while perudic soil moisture regimes favor the persistence of the short-range-order compounds.

A second mineral weathering scenario occurs under more acidic conditions when soil pH is ~<5 and organic matter is relatively abundant. Organic acids produced by the decomposition of plant materials compete with silica for the binding of aluminum (and iron). Metal-humus complexes, especially those with aluminum, are preferentially formed. These complexes tend to precipitate in the surface horizons rather than be translocated from A horizons (Dahlgren et al. 1991). The aluminum is toxic to many microorganisms, protecting the organic fraction from microbial decomposition, and

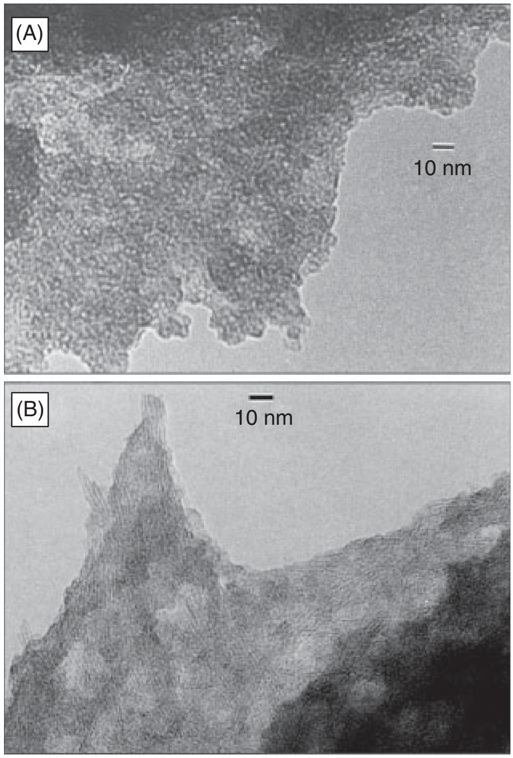

Figure 9.4. Micrographs of (A) allophane spherules and (B) imogolite threads from an Icelandic Andisol (Haplocryand). (Courtesy Geoderma and modified from Wada et al. 1992, Clay minerals of four soils formed in eolian and tephra materials in Iceland, Geoderma 52, p. 359, with permission from Elsevier)

the aluminum-humus complexes accumulate in surface horizons. Note the similarity of this process to that discussed in Chapter 17 regarding Spodosols. These Andisols are referred to as *nonallophanic* because they are dominated by organic rather than inorganic short-range-order compounds (Figure 9.5). If 2:1 layer silicate minerals are present in the parent material, the aluminum may also precipitate as hydroxy-interlayer islands. Under these conditions, desilication is a dominant process in the surface and upper subsurface horizons. Opaline silica accumulations are common, especially in semiarid climates.

Andisols generally have higher organic matter contents than do other mineral soils in similar environments. Many researchers have demonstrated positive correlations between organic matter and allophane, imogolite, and ferrihydrite. Interactions between these components appear to result in organic matter stabilization (Dahlgren et al. 2004). Organic matter may sorb to these short-range-order minerals, thereby decreasing mineralization rates. Iron and aluminum are also able to bind with humic substances, creating complexes that are very resistant to degradation and leaching (Hiradate 2004). Active aluminum associated with Al-humus complexes can be toxic to many microorganisms, thereby inhibiting decomposition and increasing the residence time of the organic fraction.

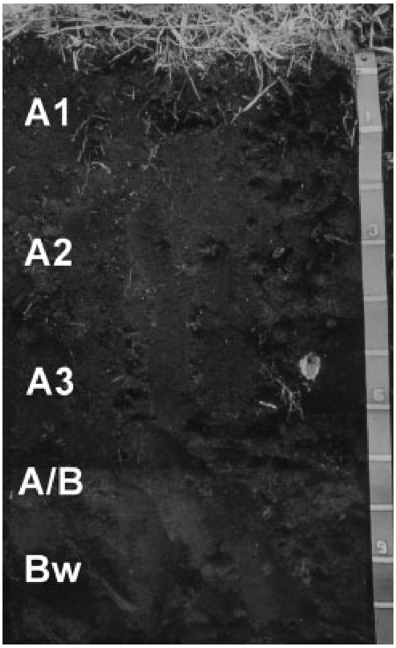

Figure 9.5. Melanudand from Costa Rica. Soil has a melanic epipedon and is an example of a nonallophanic Andisol in which organo-metal compounds dominate the colloidal fraction. Scale divisions are decimeters. For color detail, please see color plate section.

Andisols may be pedogenically linked with soils of other orders. Soils formed in fresh, unweathered volcanic ejecta have not undergone sufficient weathering to develop andic properties and are classified as Entisols. Dahlgren et al. (1997) estimated that approximately 200–300 years of weathering are required for silicic tephra to attain andic soil properties under cryic-udic conditions. Other pedogenic linkages may also occur. In cool, moist environments, Fe and Al derived from weathering of volcanic ash can form organometallic complexes that may then be translocated to form spodic horizons. Spodosols in tephra deposits have been reported in Japan (Shoji et al. 1988), New Zealand (Parfitt and Saigusa 1985), the Pacific Northwest region of the United States (Dahlgren and Ugolini 1991; McDaniel et al. 1993), and Alaska (Ping et al. 1989). Under warm, moist conditions and without significant additions of fresh volcanic ejecta, Andisols may eventually develop into soils of other orders as meta-stable weathering products are transformed into crystalline minerals. As an example, both Ultisols and Oxisols have formed from Andisols under humid, tropical conditions in Costa Rica (Martini 1976; Nieuwenhuyse et al. 2000) and Hawaii (Chadwick et al. 1999). Transformation of Andisols to Inceptisols and Alfisols has also been reported in the literature (Dahlgren et al. 2004).

Selected properties of representative Andisols are presented in Table 9.1. The Typic Vitrixerand has formed in 7,700-year-old tephra under xeric/frigid regimes, and represents a relatively early stage of Andisol development. In contrast, the Acrudoxic

Table 9.1. Selected properties of Andisols. Data for the Bonner series (Pedon no. 78P0553) and Kaiwiki series (Pedon no. 07N0375) are from the USDA-NRCS National Cooperative Soil Survey Soil Characterization Database (Soil Survey Staff 2010b). Bonner soils have a xeric soil moisture and frigid soil temperature regime; Kaiwiki soils are udic and isomesic

Horizon	Depth	Bulk density moist	Bulk density oven dry	Sand	Silt	Clay	1,500 kPa H_2O retention moist	1,500 kPa H_2O retention air dry	pH H_2O	pH NaF	C	Al + 1/2 Fe [b]	New Zealand P retention	CEC pH7	ECEC[c]
	cm	g cm^{-3}		%			%				%			cmol(+) kg^{-1}	
Bonner series (Typic Vitrixerand)—Idaho															
A	0–4	0.68	0.75	30.6	63.8	5.6	–[a]	20.3	–	9.5	13.49	1.14	38	36.9	–
Bw1	4–20	–	–	38.1	58.0	3.9	–	10.0	6.0	10.6	2.54	2.31	66	18.9	–
Bw2	20–48	0.94	0.96	39.3	59.0	1.7	–	9.9	6.1	10.6	1.34	2.32	58	13.4	–
Bw3	48–69	–	–	52.6	44.1	3.3	–	6.8	6.1	10.4	0.63	1.61	42	9.6	–
Bw4	69–89	0.80	0.80	52.7	44.8	2.5	–	6.8	6.2	10.3	0.59	1.24	37	9.4	–
2C	89–152	–	–	86.6	9.8	1.6	–	2.2	6.0	7.6	0.08	0.16	30	4.1	–
Kaiwiki series (Acrudoxic Hydrudand)—Hawaii															
A	0–16	0.78	1.09	15.3	33.5	51.2	68.1	36.2	5.2	9.3	11.68	6.12	70	46.8	4.3
Bw1	16–46	0.69	1.03	13.0	37.9	49.1	176.8	27.4	5.6	10.4	9.05	10.38	92	50.7	0.9
Bw2	46–85	0.76	1.07	11.0	42.9	46.1	149.5	24.2	5.5	10.6	6.68	8.09	92	40.7	1.0
Bw3	85–120	0.71	0.99	10.6	33.7	55.7	251.6	27.5	5.6	10.8	8.36	9.13	94	50.4	–

[a] not determined.
[b] extracted using acid ammonium oxalate.
[c] effective cation exchange capacity = bases + Al.

Hydrudand has weathered extensively under udic/isomesic regimes as evidenced by the higher measured clay contents and 1,500 kPa water retention.

It is important to recognize that not all Andisols form from volcanic parent materials. The nonvolcanic Andisols generally occur under cool, humid climates, wherein rapid mineral weathering produces abundant aluminum, which forms complexes with large amounts of organic matter in surface horizons. The Lytell series in coastal mountains of Washington state is an example of an Andisol that has formed in siltstone or fine-grained sandstone dominated by quartz, feldspar, mica, and phyllosilicate clays rather than volcanic materials (Hunter et al. 1987). High rainfall (~2,500 mm of annual precipitation), an isomesic soil temperature regime, and abundant organic acids result in the formation of Al- and Fe-humus complexes and hydrous Fe oxides of varying crystallinity (Hunter et al. 1987); this mineralogical suite is responsible for the observed andic soil properties and classification as a Typic Fulvudand (Soil Survey Division 2010).

Uses of Andisols

The uses of Andisols are limited by extremely cold conditions on high volcanic peaks and at high latitudes and by dryness in arid regions. Most Andisols are on steep slopes, thus limiting the use of much mechanization in farming operations. Many Andisols in cool, mountainous terrain support commercial timber harvest operations. Andisols of the inland mountainous areas of Washington, Idaho, and Oregon support the most productive forests found within the region (Kimsey et al. 2008). Much of this productivity is attributable to increased water-holding capacity associated with tephra mantles in these seasonally dry landscapes (McDaniel et al. 2005).

Where climatic conditions are favorable, Andisols support dense human populations. Despite having the least areal extent of the 12 soil orders, Andisols are estimated to support as much as 10% of the earth's population (Ping 2000). Most notable among these areas are parts of the Rift valley in east Africa, the Andean chain in South America, Central America, Japan, and the island of Java in Indonesia. Andisols generally have low bulk density and are easy to till. Being composed of relatively unweathered materials, they can be very fertile, with the fertility being related to the chemistry of the volcanic ejecta. Coolness, related to altitude, favors a low rate of disease and in the free-draining Andisols, there is little stagnant water to promote insect reproduction. Alluvial valleys in Andisol areas are often very fertile, although the soils may be mixed with other material and not classified as Andisols.

There can also be significant limitations to plant growth in Andisols. Siliceous tephras in combination with high rainfall and leaching, as on the North Island of New Zealand, give rise to low-fertility Andisols (Lowe and Palmer 2005). Most Andisols are acid, and in cases of very high acidity, high levels of exchangeable Al can inhibit plant growth. This problem is most common in nonallophanic Andisols (Dahlgren et al. 2004) but can be mitigated with additions of lime (Takahashi et al.

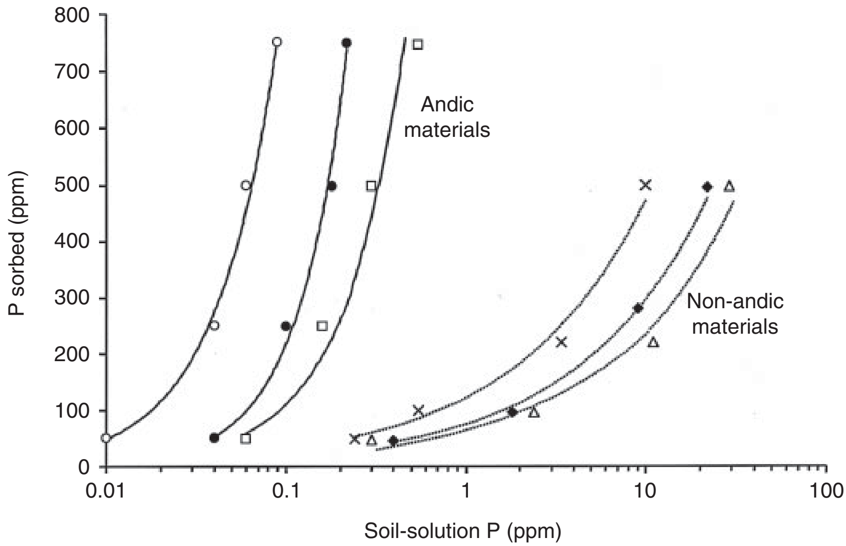

Figure 9.6. Relationship between sorbed P and soil-solution P in andic and non-andic materials in a soil profile from northern Idaho. Data points for each curve represent additions of 5, 25, 50, and 75 ppm of P. Andic materials fix virtually all of the added P and are able to maintain only low concentrations of P in soil solution. (Data from Jones et al. 1979)

2006). Allophanic Andisols are converted to nonallophanic Andisols in as little as 30 years with the establishment of bracken fern (*Pteridium aquilinum*) vegetation on clear-cut sites in northern Idaho (Johnson-Maynard et al. 1997). This shift is brought about by a substantial increase in belowground organic carbon and an accompanying increase in active aluminum associated with Al-humus complexes; the increase in active aluminum appears to be associated with poor timber regeneration.

A major limitation to improvement of Andisols for agricultural use is their tendency to strongly sorb or "fix" phosphate in a plant-unavailable form (Fox 1980). Phosphate retention results in very low soil-solution concentrations of phosphate (Figure 9.6). The phosphate appears to bind to Al-OH and Fe-OH groups via ligand exchange (Harsh et al. 2002; Schwertmann and Taylor 1989).

Because it is such an important characteristic of Andisols, high-phosphate retention is used as a criterion for defining andic soil properties in *Soil Taxonomy* (Soil Survey Staff 2010a). The highest P fixation is found in those soils that are fine grained and have relatively high Al:Si ratios. Soil material with the greatest potential to fix P can be identified by pH value >10.6 in 1N NaF (Alvarado and Buol 1985). Because large quantities of aluminum are present in both the organometallic and aluminosilicate short-range-order compounds in many Andisols, it is often not practical to overcome the fixation capacity by fertilizer application. Unique methods of P fertilization, such as adding large pellets of superphosphate to the soil or soaking potato tubers in phosphate solution have been known to work in such soils.

Most Andisols are stable and resist water erosion because of rapid infiltration and relatively high permeability. However, physical disturbance of Andisols may significantly reduce permeability and increase susceptibility to water erosion. Timber-harvesting traffic on skid trails in western Montana reduced infiltration by as much as 81% in volcanic ash-mantled soils (Cullen et al. 1991). When dry, Andisols are susceptible to wind erosion (Warkentin and Maeda 1980) and are often very dusty when trafficked for agriculture, forestry, or recreation. Work by Arnalds et al. (2001) has shown that andic materials are susceptible to wind erosion at relatively low wind velocities that are frequently exceeded in Iceland, contributing to extensive degradation of the country's highlands.

Low bulk density, poor compactability, and large changes in cohesion and friction angle on drying of Andisols require careful consideration for engineering projects (Warkentin 1985). Engineering uses of some Andisols may also be limited by a property known as *thixotropy*. Thixotropy refers to a reversible gel-sol transformation in which applied pressure causes the soil to suddenly liquify and flow (Nanzyo et al. 1993; Neall 2006). Upon release of pressure, the soil reverts to a solid state. This property is more pronounced in hydrous families of Andisols that contain large quantities of poorly crystalline weathering products, such as the Hydrudand for which data are presented in Table 9.1. The 1,500 kPa water retention for the B horizons ranges from 149 to more than 250% on undried samples. Such Andisols should be dried prior to use for many engineering purposes, because this will result in irreversible collapse of allophane spherules and greater strength (Neall 2006). Note that 1,500 kPa water retention values in Table 9.1 are substantially less for air-dried samples than for moist samples.

Classification of Andisols

The basic concept of Andisols centers on soils formed from volcanic materials that have weathered enough to produce short-range-order aluminosilicate and organometallic compounds by in situ transformation but that have not weathered to the point where crystalline materials predominate or where significant illuviation has occurred. Acid-oxalate-extractable aluminum and iron are used to estimate the quantities of short-range-order compounds that have formed from the weathering of volcanic glass, and this serves as a quantitative basis for defining andic soil properties in *Soil Taxonomy* (See Chapter 2.) Andic soil properties can be qualitatively identified using an NaF field test (Fieldes and Perrott 1966). Soil-NaF pH values in excess of 9.4–9.5 indicate the presence of allophane and/or metal-humus complexes (Soil Survey Staff 1996; IUSS Working Group WRB 2006).

In the key to soil orders in *Soil Taxonomy*, Andisols follow Histosols, which specifically exclude soils with andic properties, and Spodosols, which exclude soils with andic properties unless they have an albic horizon. Andisols have andic soil properties in 60% or more of the thickness of soil material within 60 cm of the mineral soil surface, or of the top of an organic layer with andic properties. The andic soil

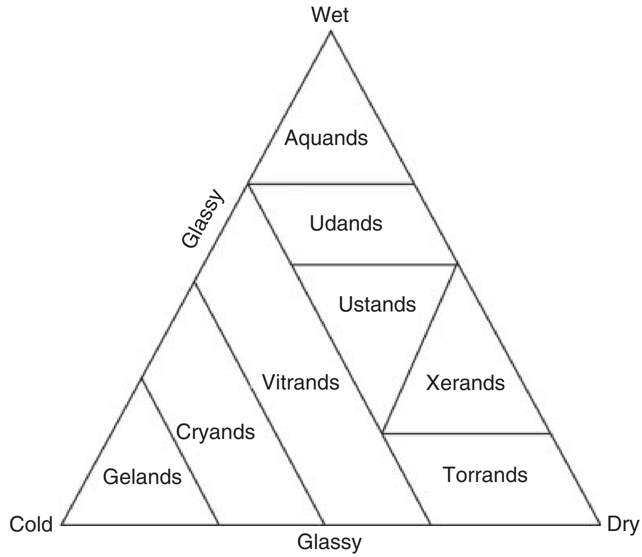

Figure 9.7. Diagram showing some relationships among suborders of Andisols.

property definition includes essentially all volcanic materials ranging from slightly altered cinders or pumice deposits to extremely fine amorphous gels of allophane and/or imogolite (Parfitt 1985; Uehara 1985).

Suborders of Andisols are defined by soil moisture and temperature regimes, and by water retention characteristics as shown in Figure 9.7 and Table 9.2.

Aquands have a histic epipedon, or have aquic conditions in a layer at a depth of 40–50 cm below the mineral soil surface or from the top of an organic layer with andic properties, and have redoximorphic features in that layer. Aquands occur locally under a wide array of climatic conditions in depressions and along floodplains where water tables are at or near the soil surface for at least part of the year.

Gelands are very cold Andisols that have a gelic soil temperature regime (mean annual soil temperature <0°C at a depth of 50 cm). However, Gelands differ from Gelisols in that they do not contain permafrost within 2 meters of the soil surface. Relatively little is known about the distribution of these soils, and it is estimated that they occupy almost 62,000 km^2 of the global land area (USDA-NRCS database). They likely occur in Alaska and the Yukon in Canada, Iceland, the Kamchatka Peninsula of Russia, and possibly high-elevation parts of southern Patagonia and the Andes along the Chile-Argentina border.

Cryands are defined as Andisols with cryic soil temperature regimes (0°C < mean annual soil temperature <8°C, with cool summers). These soils are the Andisols of high latitude (Iceland, Alaska, Kamchatka Peninsula of Russia) and high altitude (northern Cascade Range and high Sierra Nevada in the United States and the Andes in South America). Vegetation is subalpine forest or alpine shrub-tundra.

Table 9.2. Suborders and Great Groups in the Andisols order

Suborder	Great Group
Aquands	Gelaquands have a gelic soil temperature regime.
	Cryaquands have a cryic soil temperature regime.
	Placaquands have, in 50% or more of the pedon, a placic horizon within 100 cm of the mineral soil surface or the top of an organic layer with andic properties.
	Duraquands have, in 75% or more of each pedon, a cemented layer that does not slake after air-drying within 100 cm of the mineral soil surface or the top of an organic layer with andic properties.
	Vitraquands have 100 kPa water retention <15% on air-dried samples and <30% on undried samples dominant in the upper 60 cm.
	Melanaquands have a melanic epipedon.
	Epiaquands have episaturation.
	Endoaquands—other Aquands (have endosaturation).
Gelands	Vitrigelands—currently includes all Gelands.
Cryands	Duricryands have the upper surface of a cemented horizon within 100 cm of the surface in 75% or more of the pedon.
	Hydrocryands have 1,500 kPa water retention <15% on air-dried samples of 100% or more throughout a total thickness of 35 cm in the upper 100 cm.
	Melanocryands have a melanic epipedon.
	Fulvicryands have an epipedon with chroma and moist value of 3 or less that meets depth, thickness, and organic carbon requirements for a melanic epipedon.
	Vitricryands have 100 kPa water retention of <15% on air-dried samples and <30% on undried samples dominant in the upper 60 cm.
	Haplocryands—other Cryands.
Torrands	Duritorrands have the upper surface of a cemented horizon within 100 cm of the surface in 75% or more of the pedon.
	Vitritorrands have 100 kPa water retention of <15% on air-dried samples and <30% on undried samples dominant in the upper 60 cm.
	Haplotorrands—other Torrands.
Xerands	Vitrixerands have 100 kPa water retention <15% on air-dried samples and <30% on undried samples dominant in the upper 60 cm.
	Melanoxerands have a melanic epipedon.
	Haploxerands—other Xerands.
Vitrands	Ustivitrands have an ustic soil moisture regime.
	Udivitrands have a udic soil moisture regime.
Ustands	Durustands have the upper surface of a cemented horizon within 100 cm of the surface in 75% or more of the pedon.
	Haplustands—other Ustands.
Udands	Placudands have, in 50% or more of the pedon, a placic horizon within 100 cm of the mineral soil surface or the top of an organic layer with andic properties.
	Durudands have the upper surface of a cemented horizon within 100 cm of the surface in 75% or more of the pedon.
	Melanudands have a melanic epipedon.

continues

Table 9.2. *Concluded.*

Suborder	Great Group
	Hydrudands have 1,500-kPa water retention of <15% on air-dried samples of 100% or more throughout a total thickness of 35 cm in the upper 100 cm.
	Fulvudands have a layer that meets the depth, thickness, and organic-carbon requirements of a melanic epipedon.
	Hapludands—other Udands.

Torrands have aridic soil moisture regimes. These soils are of limited extent globally and occur in parts of the western United States, and in the Cape Verde, Canary, and Hawaiian islands. Most are vegetated by desert shrubs. Torrands are the least extensive of the Andisol suborders, accounting for only ~1,500 km^2 of the earth's ice-free land area (USDA-NRCS database).

Xerands have xeric soil moisture regimes. These soils occur mostly in the western United States under coniferous forest in Washington, Oregon, Idaho, and California. They also are found in southern Europe, the Canary Islands, Argentina (Broquen et al. 2005), and South Australia (Lowe and Palmer 2005).

Vitrands are defined in terms of water retention characteristics rather than by moisture or temperature regime. These Andisols have 1,500-kPa water retention of less than 15% on air-dried samples and of less than 30% on undried samples, throughout at least 60% of the thickness of soil within 60 cm of the mineral soil surface or of the top of an organic layer with andic soil properties (Figure 9.5). Put in a more general way, these are glassy and ashy Andisols that have relatively low water-holding capacity. By their placement in the key of Andisol suborders, Vitrands are restricted to ustic and udic soil moisture regimes, and support a wide variety of vegetation. By recent USDA-NRCS estimates, they are the most extensive of the Andisol suborders (USDA-NRCS database). Vitrands are extensive on the North Island of New Zealand. In the United States, they occur mostly in Holocene deposits in Oregon, Washington, and Idaho. Forest canopy and litter accumulation have helped protect the mantle of volcanic ash mantle from erosion, thereby maintaining a sufficient thickness of andic materials to meet requirements of Andisols (McDaniel et al. 2005; Soil Survey Staff 2010a). Data for an example of a Vitrand from this region are presented in Table 9.1.

Ustands have ustic soil moisture regimes. Most Ustands are in intertropical regions and present on the lee side of many of the volcanic islands in the Pacific, in Central America, South America, and Africa. These soils are cultivated to a wide array of crops, most notably coffee.

Udands have udic soil moisture regimes. These soils are almost as extensive as the Vitrands and are certainly the most studied, particularly in Japan and New Zealand. This suborder represents the classic dark-colored volcanic soils with low bulk density, high phosphate-fixing capacity, and abundant aluminum and iron in short-range-order organic and inorganic compounds. In the United States, Udands occur in Hawaii (as shown in Table 9.1), and on the west side of the Cascade Range in Washington and Oregon.

Great groups are identified on the basis of a number of properties, including placic horizons, cemented layers, water retention characteristics, and characteristics of the epipedon. (See Table 9.2.) Especially noteworthy is the melanic epipedon, which is identified in great groups of the Aquands, Cryands, Xerands, and Udands. This epipedon is dominated by black Type-A humic acids that commonly are derived from graminoid roots (Figure 9.5). The melanic epipedon has a thickness of <30 cm, andic soil properties throughout, a chroma and moist value of 2 or less, at least 6% organic carbon as a weighted average and at least 4% organic carbon in all layers, and a melanic index of 1.70 or less. (See Chapter 2.) The melanic index measures the ratio of Type A humic acid to other humic acids (Leamy 1988). The melanic epipedon originally was thought to form only under grass vegetation. Subsequently, it has been shown to also occur under coniferous forest (Southard and Southard 1989) where fire may play an important role in its formation (Takahashi et al. 1994).

Subgroups of Andisols are recognized by a number of soil properties, including exchangeable aluminum, base saturation, cation exchange capacity, and presence or absence of other diagnostic surface and subsurface horizons. The relatively small number of subgroups that currently exist for Andisols will undoubtedly increase as these soils and their classification in *Soil Taxonomy* receive more attention. The presence of thin near-surface horizons with andic soil properties is a criterion for the identification of subgroups (e.g., Andic, Vitrandic) of several other orders that intergrade to the Andisols. These intergrades include soils in which surficial processes have mixed and diluted volcanic ash with other non-ash material in such a way that the resulting admixture does not exhibit sufficient andic properties to qualify as an Andisol (McDaniel and Hipple 2010).

The presence in Andisols of short-range-order compounds and glass instead of distinct clay-sized particles or silicate minerals renders conventional particle-size and mineralogy criteria essentially useless as morphological or classification criteria at the family level. Substitute names for particle-size classes are based on size and composition of volcanic fragments (e.g., pumiceous and cindery) and on water-retention characteristics (e.g., ashy, medial, and hydrous). Family mineralogy of Andisols is typically one of four classes (amorphic, ferrihydritic, glassy, and mixed) that focus on the properties of the short-range-order compounds and volcanic glass. See Chapter 7 for a more detailed discussion of family criteria.

Perspective

Soils formed in volcanic ejecta share many features that distinguish them from most other soils (Arnold 1985). Their geographic distribution is linked closely to Holocene volcanism. Andisols are dominated by short-range-order inorganic and organic compounds, and are distinguished from Spodosols, which share many similar properties, by the processes of in situ mineral and organic transformations and lack of extensive illuviation of these compounds. These soils are used extensively for agricultural production in warmer, more humid climates. Andisols are generally characterized by a combination of low bulk density, high phosphate sorption capacity, relatively high amounts of aluminum and iron that can be extracted from short-range-order or nanominerals, and volcanic glass—collectively referred to as andic soil properties. Suborders, great groups, and families are identified largely based on soil moisture and temperature regimes, water retention characteristics, and characteristics of the epipedon. The establishment of the Andisol order in 1990 reflects the continued growth of soil science and demonstrates the ability of *Soil Taxonomy* to change as we learn more about soils.

Aridisols: Soils of Dry Regions

10

Aridisols occur in both cool temperate deserts (between latitudes 35 and 55) and warm deserts at lower latitudes (Cameron 1969). Soils with permanently frozen subsoils in dry polar regions (Campbell and Claridge 1990) are classified in the Gelisol order (Chapter 12).

Setting

Deserts occupy about one-third of the areas of Africa and Australia, 11% of Asia, and only about 8% of the Americas. Only a small proportion of the deserts of the planet consist of barren sand dunes and rock land, movie sets for the *Arabian Nights* notwithstanding. Rather, most of these areas are surprisingly well vegetated with scattered plants (Figure 10.1), the root systems of which extend considerable distances both laterally and vertically from each plant. Various species of cactus (*Cactaceae*), mesquite (*Prosopis*), creosotebush (*Larrea*), yucca (*Yucca*), sagebrush (*Artemisia*), shadscale (*Atriplex*), hopsage (*Grayia*), and muhlygrass (*Muhlenbergia*) are common. Microbial populations are low. Low carbon to nitrogen ratios in the sparse soil organic matter are probably attained by action of nitrifying bacteria and/or nitrogen-fixing blue-green algae that form a crust on some of these soils (Mayland et al. 1966; Evans and Johansen 1999; Johnson et al. 2007).

Arid regions occupy 36% of the earth's land surface on the basis of climate and 35% on the basis of natural vegetation (Shantz 1956). Aridisols do not conform to all the parameters of either climatic or vegetative zones (Buol 1965; Dregne 1976; Southard 2000). The moisture control section of these soils is dry in all parts more than 50% of most years and not moist in any part as much as 90 consecutive days when the soils are warm enough (>8°C) for plant growth. In an aridic soil moisture regime, potential evapotranspiration greatly exceeds precipitation during most of the year, and in most years no water percolates through the soil. Figure 10.2 (Buol 1964) shows a characteristic aridic water balance where about 1.3 cm (0.5 in.) of water is recharged (R) into the soils during December, January, and February. This stored water is utilized (U) during March, and the soils are deficient (D) in water throughout most of the year. Even though leaching is limited, many Aridisols have morphologically distinct horizons due to the concentration of the limited water in a relatively small volume of soil.

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

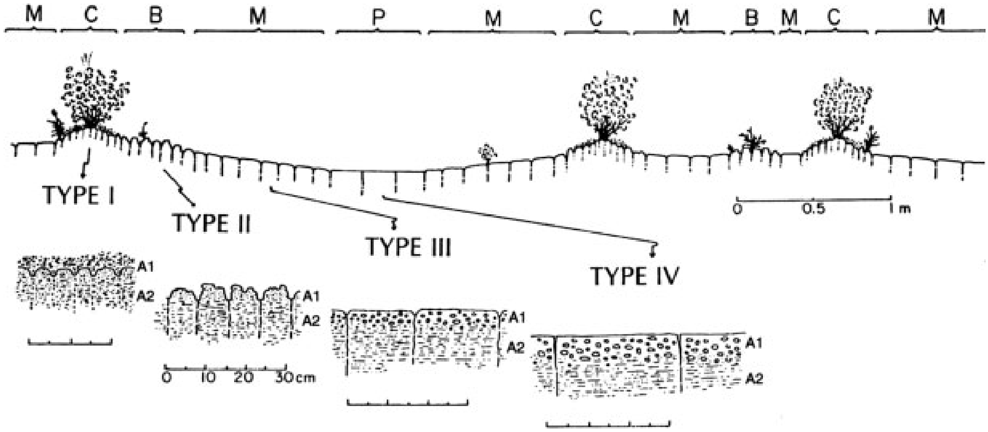

Figure 10.1. Schematic cross-section diagrams of the microtopographic positions and associated surfacesoil morphological types of loess-mantled Argids of the Humboldt loess belt of Nevada, with big sagebrush (*Artemisia tridentata* var. *wyomingensis*) plant communities. Positions are C = coppice, B = coppice bench, M = intercoppice microplain, P = playette. M and P can be wider than shown; several coppices may be linked together. *Vertical lines* below the soil surface show sides of surface crust polygons (A1 horizon) that extend into the underlying A2 horizon with compound weak prismatic and moderate platy structure. Type I is weakly crusted and litter covered. Type II is pinnacled. Only Types III and IV are significantly crusted. *Circles* indicate vesicular pores in the A1 horizons of Types III and IV. (Courtesy, Society for Range Management, Denver, Colorado, and modified from Eckert et al. 1978)

Soil-vegetation patterns of arid regions are recognized at many scales. Satellite imagery shows clearly demarcated, interspersed bright areas, which are sandy and saltation-prone under the impact of wind, and relatively dark areas, which are stable, crusted, and "immune" to wind erosion (Otterman and Gornitz 1983). Figure 10.1 shows fine-scale patterns of vegetation and soils as observed in Nevada (Eckert et al. 1986). Patterns of runoff and run-on are intricate and important in determining distribution and condition of desert vegetation (Schlesinger and Jones 1984; McAuliffe 1999), nutrient cycling, and leaching of soluble salts (Reid et al. 1993; Wood et al. 2005; Graham et al. 2008). Playas and other utterly barren sites are considered by some soil scientists to be without soils.

Aridisols occupy about 12% of the global land area. Only about one-third of the arid terrains are occupied by Aridisols. Other soils present are torric, ustic, and xeric great groups of Entisols, Mollisols, and other orders (Figures 10.3 and 10.4). On a transect from arid to adjacent semiarid areas (left to right, Figure 10.3), ochric epipedons darken, with increasing organic matter content (as precipitation increases) and merge into mollic epipedons, preferentially in fine-textured and/or calcareous parent materials. Alfisols appear on adjacent forestlands and on some grassland. Vertisols are present on bodies of lithogenic or pedogenic smectite clays.

Evidence of leaching below the average depth of water storage is often observed in Aridisols and attributed to pluvial times associated with Quaternary climatic

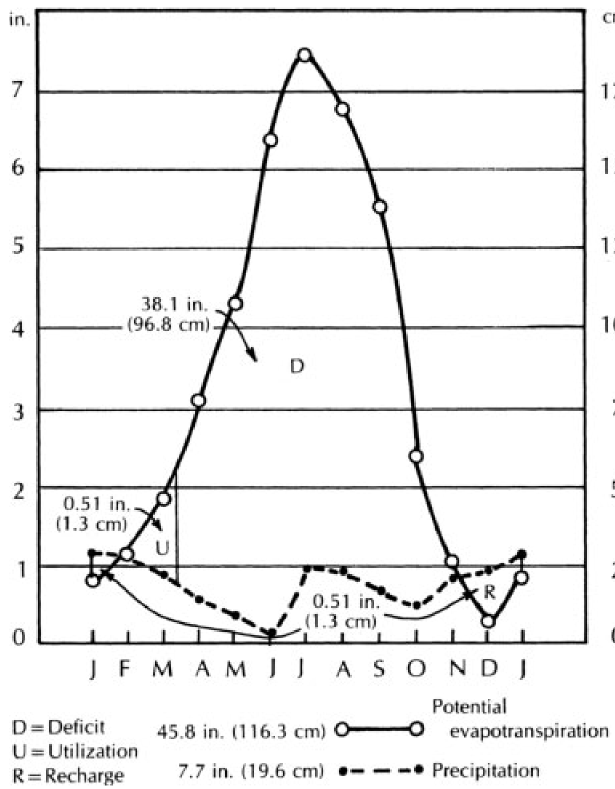

Figure 10.2. Aridic soil moisture regime for soils at Phoenix, Arizona. Note that the soil moisture control section is dry in all parts more than half of the time in an average year.

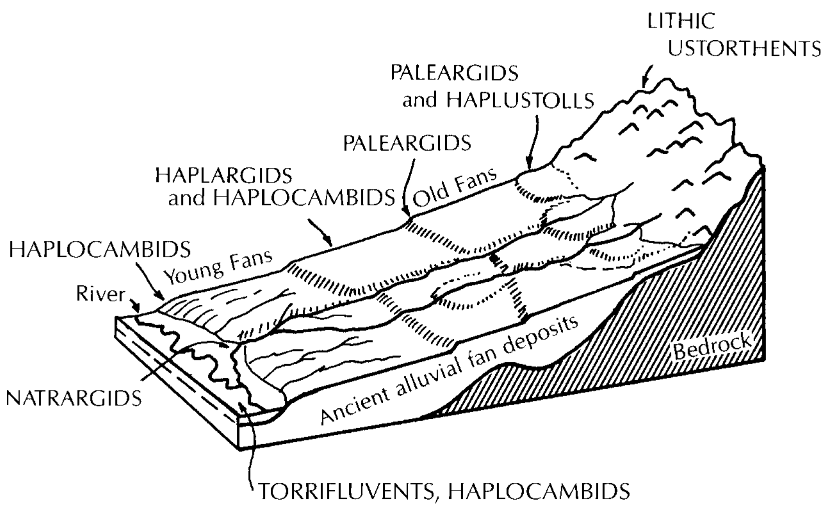

Figure 10.3. Block diagram showing positions of some major kinds of Aridisols and their associates.

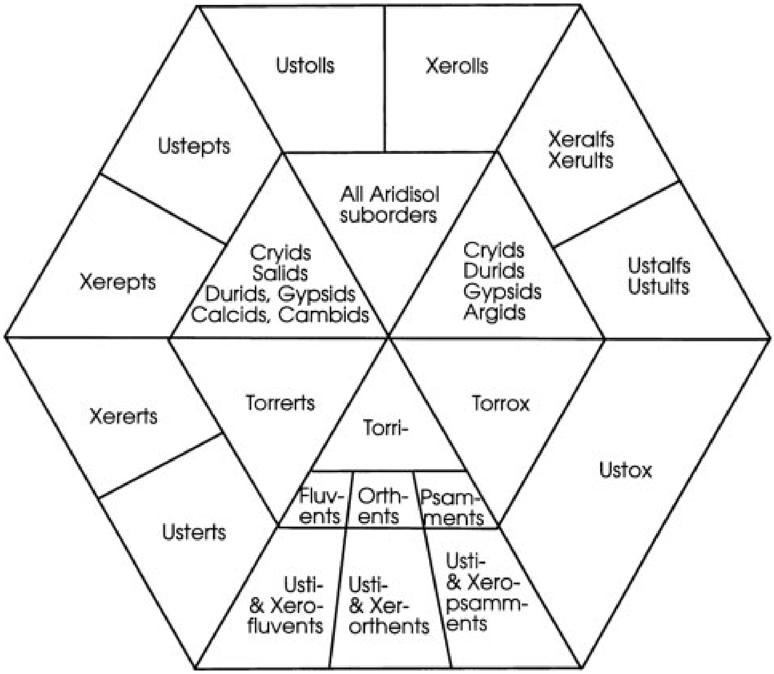

Figure 10.4. Relationships between suborders and great groups of arid-land and semiarid-land soils. (After Flach and Smith 1969)

changes (Smith 1965; Wells et al. 1987). Another explanation is that the typically erratic distribution of the rainfall causes occasional periods of relatively high precipitation in the winter months, during which deeper leaching takes place. In this case, the depth of leaching may reflect the rainfall of the extreme years, rather than that of average years.

Pedogenic Processes

Pedogenic processes in desert regions have produced numerous soil features. Of special interest are (1) physical surface crusts, (2) biological soil crusts, (3) vesicular horizons, (4) desert pavement, (5) cambic horizons, (6) argillic and natric horizons, (7) pedogenic calcium carbonate accumulations (calcic and petrocalcic horizons), (8) duripans, and (9) gypsic and salic horizons (Nettleton and Peterson 1983).

A physical soil crust (Figure 10.5A), widespread in arid regions, is a thin surficial layer of uncemented fine earth that is coherent when dry and can be broken free from underlying soil material. Its strength arises from physical cohesion rather than biological influences. The crust is generally less than 10- to 20-mm thick and is often massive. Such crusts may grade into vesicular horizons as described below. Where present, physical crusts can impede infiltration and seedling germination.

(A) (B)

Figure 10.5. Examples of (A) physical and (B) biological crusts on Aridisols. In each case the crust is about 1-cm thick. In (A), rain splash and dispersion of sodium-saturated, smectite-rich silty clay soil material destroyed its original structure and produced a platy crust (Carrizo Plain, California). In (B), a biological soil crust composed mainly of cyanolichen stabilizes the soil surface and fixes nitrogen (Clark Mountains piedmont, Mojave Desert, California). (Photo by Nicole Pietrasiak)

Biological soil crusts (formerly called “microbiotic” or “cryptogamic crusts”) (Figure 10.5B) are dominant features of many arid ecosystems. They are actually a crust of living organisms—intricately intertwined lichens, cyanobacteria, mosses, and algae—growing on the soil surface. Low water availability in arid environments results in discontinuous plant cover with soil exposed to direct sunlight in the interspaces. Biological soil crusts often form a continuous photosynthetic layer on the soil surface in these openings. These crusts are biologically dormant during the prevailing dry conditions but are almost immediately activated by rainfall. Biological crusts stabilize the soil surface against wind and water erosion, trap windblown silt and clay particles, and fix nitrogen. Various studies have shown that these crusts can increase, decrease, or have no effect on infiltration. The effect of biological soil crusts on hydrologic behavior, such as infiltration and runoff, seems to vary with the dominant biotic component and local site characteristics (Evans and Johansen 1999).

Vesicular horizons (Figure 10.1. and Figure 10.6A) are another very common surface feature of arid zone soils worldwide. These horizons form in fine-textured eolian material at the surface or, commonly, immediately below desert pavement. They are commonly labeled “Av horizons,” but this is not an accepted designation in the USDA system (Springer 1958; Anderson et al. 2002). The more strongly developed vesicular horizons have a compound structure consisting of prismatic or columnar units, often on the order of 10 cm in diameter, which in turn have platy structure and vesicular pores (Figure 10.6b). Repeated wetting and drying cycles, often associated with intense summer thunderstorms, produce these features. As the soil wets, escaping gases form bubbles (vesicular pores) that are preserved when the soil dries (Springer 1958). With repeated wetting and drying, the bubbles enlarge, coalesce, and eventually collapse. The planes of weakness established by the collapsed bubbles yield the characteristic platy structure (Miller 1971). The coarse prismatic structure results

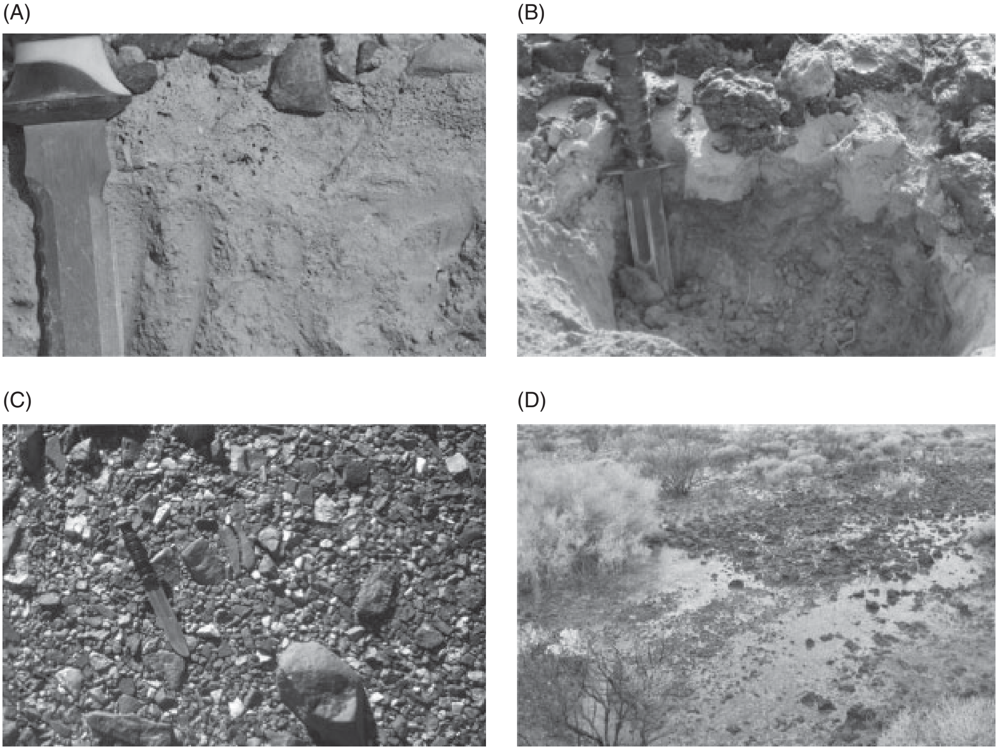

Figure 10.6. Examples of desert pavement and associated vesicular horizons in the Mojave Desert, California: (A) a 3-cm-thick vesicular horizon (photo by Nathan Bailey), (B) desert pavement overlying an 8-cm-thick vesicular horizon with columnar structure, (C) closely interlocking rock fragments make up a desert pavement (knife is 30 cm long), and (D) water ponding on desert pavement during a rainstorm.

from relatively slow desiccation of uniformly textured soil material (Chadwick and Graham 2000). When wet, the prismatic structural units swell, closing the cracks between them. This, along with the platy structure and lack of pore continuity, result in low permeability of water (Young et al. 2004), in contrast to the rapid infiltration that happens in uncrusted coppice sand dunes (Figure 10.1) and vegetated areas (Gile 1966a; Eckert et al. 1979; Reid et al. 1993). The low infiltration rates typical of desert pavement/vesicular horizon surface conditions cause the subsoil accumulation of high levels of eolian-derived salts, including nitrate, while adjacent soils under shrubs are well-leached and nonsaline (Graham et al. 2008).

The complexity of processes and morphology in strongly developed vesicular horizons is described by Anderson et al. (2002). The columnar peds of a 10-cm-thick vesicular horizon in a Typic Natrargid (Figure 10.6b) were found to have up to 40% clay and 12% pedogenic calcite in their interiors, whereas the material adhering to the ped sides contained less than 7% clay and 2% calcite. Desert dust enriched in clay and

Figure 10.7. Photograph of a fine, smectitic, thermic Natric Petroaragid near Laguna Chapala, Baja California. Note the vesicular horizon (0–7 cm) under a desert pavement, the natric horizon (7–70 cm), and calcic horizon (15–70 cm). Laminar petrocalcic material caps the boulders in the subsoil. For color detail, please see color plate section.

calcite is trapped by the rough surface of the overlying desert pavement. It is eluviated down the cracks between the columns, and then is carried laterally by infiltrating water into ped interiors via cracks between platy structural units. Argillans and siltans line the platy surfaces in the lower part of the columns. Several lines of evidence suggest that these clay- and calcite-rich vesicular layers experienced much of their development within the last 5,000 years, when desiccation of local playas provided a source of abundant eolian materials.

Many Aridisols that developed in stony material have surface pebble layers (Figure 10.6b, 10.6c, Figure 10.7), variously called desert pavement, stone pavement, gibber, gobi, sai, hammada, and reg. Desert pavements form on stable geomorphic surfaces, such as alluvial fans and lava flows. They are composed of pebbles, cobbles, and stones derived from the underlying alluvium or bedrock. A number of mechanisms have been proposed for the formation of the pavement, including concentration of clasts at the surface by wind and water erosion and upward migration of clasts through the soil by shrink-swell heaving and other physical processes (Cooke 1970). Strong evidence has now accumulated showing that most pavements form on accreting landscapes as rock fragments on the original surface are rafted upward by infiltrating eolian material (Wells et al. 1985; McFadden et al. 1987). The pavement itself serves as a dust trap (Yaalon and Ganor 1973; Gile 1975a; Peterson 1977). Fine particles lodge between clasts and are later eluviated down-profile during rains, becoming part of the underlying vesicular layer, as described above. Thus, desert pavement and the underlying vesicular layer function together as a unit to trap dust, incorporate it into the soil, and build the surface upward. Surficial weathering, including fracturing by heat stress and salt crystallization (Cooke 1970), splits vulnerable clasts, and the fragments become part of the pavement (McFadden et al. 2005). Over tens of thousands of years these processes lead to a closely interlocked mosaic of clasts that nearly completely covers the surface. The effectiveness of the pavement as a dust trap decreases as the surface fragments become truncated and more interlocked. At this point, the surface neither retains much new eolian material, nor loses much material to erosion, unless disturbed. Eventually, however, landscapes with highly developed desert pavement will self-destruct, as shown for pavement landscapes older than 700,000 years in the Mojave Desert of California (Wells et al. 1985). The tightly interlocked pavement and the underlying soil enriched in pedogenic clay and calcite combine to yield very low infiltration rates so that rainwater ponds on the surface and runs off (Figure 10.6d). Concentrated surface runoff from heavy rains causes erosion and the development of extensive drainage networks on the once-stable pavement and soil.

Exposed surfaces of desert pavement clasts develop a black coating, 10–500 μm thick, known as desert varnish. This varnish is composed of eolian-deposited silicate clays and microbially precipitated iron and manganese oxyhydroxides (Dorn and Oberlander 1981). Desert varnish coats not only pavements but also rock outcrops and stone archaeological features. The strong, black pigmenting effect of manganese oxides overrides other influences to give the varnish its characteristic color. The undersides of desert pavement stones are reddish (Helms et al. 2003), suggesting that this microenvironment experiences redox potentials low enough to prevent manganese oxide accumulation, yet high enough to allow precipitation of the reddish-pigmenting iron oxides.

Cambic horizons (Gile 1966b) are somewhat altered subsurface horizons of stable landscapes; mixing has obliterated stratification or rock structure; chroma may be relatively high; soil structure may be present (prismatic, blocky); a slight accumulation of clay may be evident; an underlying Bk horizon may indicate that carbonates have been eluviated from the overlying Bw horizon. Most Cambids developed in low-carbonate parent materials are leached free of carbonates in the cambic horizon. In highly calcareous parent materials, evidence of carbonate removal may take the form of carbonate coatings on undersides of pebbles in the cambic horizon, which overlies a horizon of distinct carbonate accumulation.

Argillic horizons, ranging in texture from sand to clay, commonly range in thickness from 7.5–75 cm, and begin at shallow depths (4–25 cm) below the soil surface (Figure 10.8). Smith and Buol (1968) reported evidence for both in situ weathering and illuviation of clay in Bt horizons of two Argids in Arizona. Because clay flocculates in the presence of carbonates, we may assume that accumulation of illuvial clay (largely clay-size mica: see Buol and Yesilsoy 1964; Paredes and Buol 1981; and smectite: see Graham and Franco-Vizcaíno 1992; Eghbal and Southard 1993a; Boettinger and Southard 1995) in argillic horizons did not start until the upper solum was more or less free of carbonates. The upper boundary of these horizons is often

Figure 10.8. Photograph of a coarse-loamy, mixed, superactive, thermic Typic Petroargid near Las Cruces, New Mexico. Note the platy structure of the petrocalcic horizon beginning at the 150-cm depth. For color detail, please see color plate section.

abrupt. The lower boundary may be engulfed by younger accumulations of $CaCO_3$, principally below the argillic horizons (Figure 10.7).

Most Aridisols with argillic horizons occur on earliest Holocene and older landscapes. Horizons of clay accumulation may be formed in 2,000 to 6,600 years where >15% saturation of exchangeable sodium, or of sodium and magnesium together, permits dispersion and eluviation of calcareous clay (Alexander and Nettleton 1977; Peterson 1980). The Na-enriched argillic horizon, with prismatic or columnar structure and sufficient exchangeable Na, is called a natric horizon (Figure 10.6b, Figure 10.7). Some of the argillic horizons on old landscapes have been converted to natric horizons in recent times, by additions of aerosolic sodium salts (Figure 10.7).

The requirement of identifiable argillans has been waived in defining argillic horizons of Aridisols with appreciable shrink-swell activity (usually clayey and smectite-rich). Clay skins are usually absent from ped faces and walls of voids in these soils. Theoretically, argillans once existed but were disrupted by shrink–swell of the clayey peds. Oriented clay bodies have been observed inside peds and are thought to be remnants of shattered argillans. Argids with abundant coarse skeletal fragments retain oriented clay as coatings on and within coarse sand particles and rock fragments, which, unlike clayey peds, do not shrink and swell and fracture argillans. Likewise, argillans are preserved within fractures in underlying bedrock, as reported for Argids in Baja California (Graham and Franco-Vizcaíno 1992). Remnants of argillic horizons are preserved within “pipes” in petrocalcic horizons of Paleargids

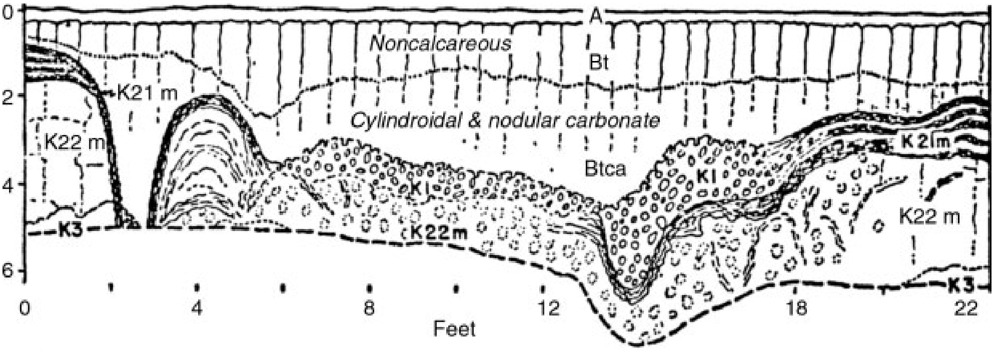

Figure 10.9. Scale diagram of horizons associated with a large pipe penetrating the thick K horizon (Bkm) of a Cacique soil on the La Mesa geomorphic surface. Scale is in feet. (Adapted from Gile et al. 1966, ©Wilkins and Williams, with permission)

(as indicated in lower left, Figure 10.9). The pipes are lined with dense, laminar $CaCO_3$ and the argillic horizon material within them has argillans on ped faces that have been kept leached of carbonates by the extra discharge of runoff waters through the pipes.

Horizons of calcium carbonate accumulation commonly lie below argillic and cambic horizons of Aridisols (Figures 10.7 and 10.8), although A horizons of many of these Argids and Cambids are weakly calcareous due to influx of calcium-containing dust during the more arid Holocene. Pedogenic calcium carbonate is almost always calcite, but dolomite has been reported (Capo et al. 2000). Figure 10.10 shows a sequence (left to right) of these calcium carbonate-enriched horizons in both gravel-free and gravelly materials, from weakly developed (Bk horizon; stage I) through moderately developed (calcic horizons: Bk horizon, stage II, and Bkk horizon, stage III) to strongly developed petrocalcic horizons (stage IV). Very old petrocalcic horizons develop distinctive features such as thick laminar layers and platy structure (stage V) and brecciation and recementation (stage VI) (Birkeland 1984). Marbut (1928b) defined the concept "Pedocal" (now obsolete) for soils with horizons enriched in calcium carbonate relative to the horizons both above and below them. Geologists have referred to the more strongly developed of these horizons (stage III and higher) as caliche or calcrete. Palygorskite and sepiolite are common clay minerals in stage IV and higher petrocalcic horizons (Monger and Daugherty 1991; Graham and Franco-Vizcaíno 1992), where the alkaline, silica- and magnesium-rich environment favors their formation (Singer 1989).

The soil shown in Figure 10.7 has a stage II carbonate accumulation in the 15- to 35-cm depth range and a stage III horizon from 35 to 70 cm. The soil in Figure 10.8 has a stage IV petrocalcic horizon beginning at the 150-cm depth. The platy structure apparent at this depth is common in petrocalcic horizons.

A master K horizon (K from German *Kalk*, lime) characterized by "K-fabric" (fine-grained, authigenic carbonate coatings and fillings) was proposed by Gile et al.

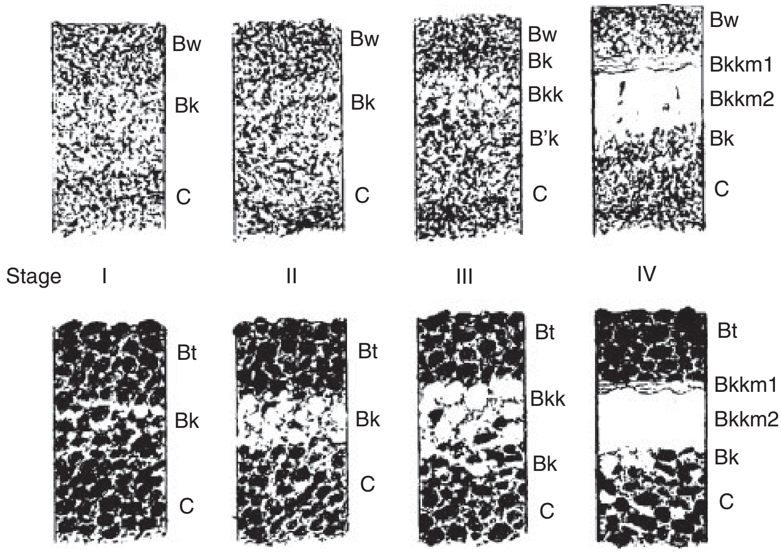

Figure 10.10. Sequential diagrams illustrating a four-stage progressive accumulation of $CaCO_3$ in nongravelley (*top row*) and gravelly (*bottom row*) materials. (After Gile et al. 1966 and Nettleton and Peterson 1983). Accumulations begin as filaments of $CaCO_3$ (*top row, left*) and coatings on undersides of stones (*bottom row, left*). After plugging of voids is finally complete (*both rows, right*), infiltrating rainwater is perched above the plugged zone, causing resolution and reprecipitation of carbonates and enabling the laminated cap (Bkkm1) to grow upward.

(1965, 1966) but has not been incorporated into USDA soil horizon designation system (Soil Survey Staff 1975, 1998). Horizons of secondary carbonate accumulation are indicated by the master B horizon modified by the following subscripts: k indicates carbonate accumulation, kk indicates engulfment of the horizon by carbonates, and km identifies extensive cementation by carbonates. Figure 10.10 uses the USDA notations.

Whereas depths to calcic horizons (≥15% $CaCO_3$ equivalent; air-dry fragments will slake in water) increase along a transect from low to higher rainfall, petrocalcic horizons (so continuously cemented with carbonates that fragments will not slake in water) lie at depths that do not correlate with present-day rainfall patterns. Many petrocalcic horizons are relicts of older landforms and are not related to current environmental conditions, although upper, laminar layers may be the result of present-day processes of localized dissolution and reprecipitation (Harden et al. 1991; Eghbal and Southard 1993b). Features within petrocalcic horizons, such as laminar layers, brecciated zones, and carbonate pendants, have been used to reconstruct paleoenvironmental conditions dating back several million years (Brock and Buck 2009). Horizons overlying petrocalcic horizons and calcareous duripans may be composed of materials much younger than the cemented horizon itself, due to erosion and redeposition processes (Blank et al. 1998).

Pedogenic Bk horizons form by eluviation of dissolved constituents of $CaCO_3$ from upper horizons and their illuviation into lower horizons, where they precipitate (Birkeland 1974). Percolating rainwater in which carbon dioxide from respiring roots and decomposer microflora is dissolved drives this process.

$$CO_2 + H_2O \rightarrow\leftarrow H_2CO_3 \rightarrow\leftarrow H^+ + HCO_3^- \text{ (carbonic acid)}$$

Solution of carbonate in upper horizons and precipitation in lower ones is represented by the following equation:

$$\begin{array}{l}\text{(Calcification } \rightarrow\leftarrow \text{ Decalcification)}\\ Ca^{2+} + 2HCO_3^- \ \rightarrow\leftarrow\ CaCO_3 + H_2O + CO_2\end{array}$$

The release of biogenic carbon dioxide into the soil can increase the CO_2 content of soil air to as much as 100 times that of the atmosphere (Jenny 1980), although gases extracted from Aridisols in southern Nevada had CO_2 levels only up to 10 times that of the atmosphere, with the highest values during the spring (Amundson et al. 1989; Terhune and Harden 1991). As bicarbonate-bearing water percolates down from the A horizon in Aridisols, it enters subhorizons with lower CO_2 concentrations due to minimal biological activity, and lower moisture contents caused by previous evapotranspirational drying. Both of these conditions favor the precipitation of $CaCO_3$, although concentration of the soil solution by evapotranspiration is generally an overriding factor in precipitation of calcite in Aridisols.

Both morphological and experimental evidence show that $CaCO_3$ precipitation in soils is not simply an inorganic chemical reaction but is strongly influenced by microbial activity. Micromorphological examinations reveal abundant calcified biological structures, such as fungal hyphae, within calcic horizons (Monger et al. 1991a). In a laboratory experiment, calcite was found to precipitate in inoculated soils, but not in otherwise identical sterile soils (Monger et al. 1991b).

Atmospheric deposition is a major contributor of $CaCO_3$ in many Aridisols. In the vicinity of University Park, Las Cruces, New Mexico, local sources of calcareous (and perhaps locally gypsic) dust have been well documented (Gile 1975b, 1975c; Gile and Grossman 1979). Dustfall has been so ubiquitous for millennia that all but the youngest soils (Entisols) have carbonate accumulations, whether or not the parent materials are calcareous (Gile et al. 1966). An external source of carbonates is assumed in view of several considerations, above all that an unrealistically large volume of primary rocks would have to be weathered to yield the volume of carbonates present in some calcic horizons (Birkeland 1984). Throughout southern Nevada and California, particulate $CaCO_3$ accounts for 10 to 30% of the annual dustfall (1–6.6 g m^{-2} yr^{-1}) (Reheis and Kihl 1995). Even when the dust does not contain $CaCO_3$, rainwater may contribute Ca that substantially enhances the precipitation of calcite in the soil (Rabenhorst et al. 1984).

In other Aridisols, mineral weathering (primarily Ca-feldspars and hornblende) is the major source of Ca in carbonates, especially on Pleistocene-age landscapes (Boettinger and Southard 1991). Data presented by Jenny (1980) indicate that cool season rains of the xeric soil moisture regime in California have been more effective than warm summer showers of the ustic soil moisture regime in the western Great Plains in leaching carbonates from upper horizons of soils.

In tropical regions, extensive areas of Aridisols have formed in parent materials that contain few easily weatherable minerals. The absence of carbonates, either autochthonous or allochthonous, is associated with thick argillans, irrespective of moisture regimes that range from aridic to udic. Paredes and Buol (1981) observed, in such a climosequence, three Ustollic Haplargids that, like the Alfisols and Ultisols in the same sequence, had thick (up to 3-m depth) argillic horizons.

Duripans are horizons cemented primarily with opal (Flach et al. 1973; Chadwick et al. 1987a), and possibly by chalcedony and quartz (Flach et al. 1973; Smale 1973). Geologists often call these horizons silcretes and duricrusts (Jackson 1957), although some silcretes and duricrusts are largely of geologic, not pedologic, origin. Most duripans in Aridisols contain considerable amounts of calcium carbonate, up to about 70% by weight, but at least half the volume of a duripan does not slake in acid, as petrocalcic horizons do (Southard et al. 1990). Many duripans have indurated, laminar upper layers up to about 2-cm thick that overlie massive and less strongly cemented, highly calcareous material, and can be distinguished from a petrocalcic horizon only by the acid treatment. In many cases only this upper laminar layer survives the acid treatment intact. The silica cements are derived from weathering of siliceous rocks (Boettinger and Southard 1991) and sediments, including volcanic ash (Chadwick et al. 1987b) and loess (Blank and Fosberg 1991).

Gypsic and salic horizons are typical of playa basins (Driessen and Schoorl 1973) where a high water table limits leaching, and eolian salts are recycled from the playa to surrounding terrain (Eghbal et al. 1989; Hirmas 2008). Improper application of irrigation water has in places raised saline groundwater into the rooting zone of plants, interfering with their growth and forming salic and gypsic horizons.

Uses of Aridisols

Use of Aridisols for agriculture is limited chiefly by the lack of water. The soils shown in Figure 10.11 are used for range except for level areas on which irrigation is practiced. However, many Aridisols present some problem with irrigation and in any case are commonly less well situated for watering than are the associated Torrifluvents. Where irrigation of Aridisols is anticipated, it is usually necessary to level the land, but drip and pivot irrigation of Aridisols is becoming more common, reducing the need for land leveling. Where leveling is needed, calcic, petrocalcic, natric, or argillic horizons or duripans may be exposed. Only soils with internal permeability adequate for deep leaching should be selected for irrigation to avoid problems of salinization and alkalization arising from salt contained in irrigation water.

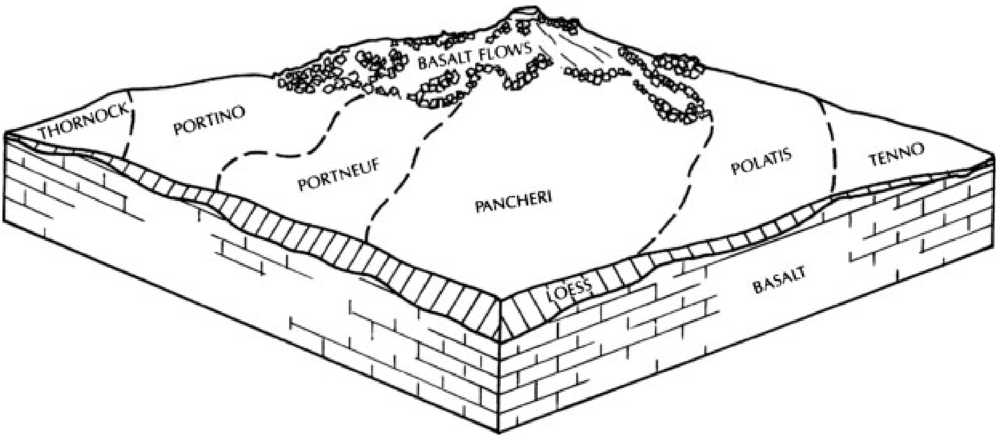

Figure 10.11. Soilscape pattern of six Aridisols. Four are coarse-silty, mixed: Portneuf soils are mesic, Durinodic Xeric Haplocalcids; Portino soils are mesic, Xeric Haplocalcids; Pancheri and Polatis are frigid, Xeric Haplocalcids. Two are loamy, mixed: Thornock soils are mesic, Lithic Xeric Haplocalcids; Tenno soils are frigid, Lithic Xeric Haplocambids. The soilscape is in the Bingham area, Idaho. (From Salzmann and Harwood 1973)

Although N content is generally low in nonsaline Aridisols, other major plant nutrient elements are often abundant, especially K from feldspars and mica. Supplies of micronutrients are usually plentiful, although they may not be available because of the high pH. Foliar applications of iron and trace elements may be necessary for satisfactory crop production.

Wind erosion, particularly when land is cultivated or otherwise disturbed, is a common problem in Aridisol landscapes where strong winds tend to predominate. Crops can be severely damaged by physical abrasion from blowing sand. Dust production has increased worldwide due to anthropogenic disturbances of arid lands, and this fine particulate matter can pose a serious human health hazard (Dahlgren et al. 1997).

Without irrigation the Aridisols are useful to only a limited extent for seasonal grazing, most of which is concentrated on the associated Torrifluvents. Few engineering problems are encountered on the Aridisols. Flash floods are a hazard along drainageways. Slow permeability of natric horizons may affect septic tank filter field performance, and petrocalcic horizons and duripans may impede excavation in building and landscaping. Large open-pit mining operations are located in Aridisol landscapes of the western United States. Stabilization and restoration of these drastically disturbed lands present severe challenges because soil moisture is so limiting (Gillis 1992). Aridisols are also used for off-road vehicle recreation and military training operations, both of which can disrupt fragile surficial features, including microbiotic crusts and desert pavement, leaving the soil susceptible to erosion (Eckert et al. 1979; Goossens and Buck 2009). Such disruption is slow to heal; tank tracks from the World War II training exercises commanded by General

Table 10.1. Suborders and great groups in the Aridisol order

Suborder	Great Groups
Cryids	Salicryids—salic horizon within 100 cm of the surface
	Petrocryids—duripan, petrocalcic, or petrogypsic within 100 cm of the surface
	Gypsicryids—gypsic horizon within 100 cm of the surface
	Argicryids—have argillic or natric horizon
	Calcicryids—calcic horizon within 100 cm of the surface
	Haplocryids—other Cryids
Salids	Aquisalids—saturated with water within 100 cm of the surface for 1 or more months in most years
	Haplosalids—other Salids
Durids	Natridurids—have a natric horizon above the duripan
	Argidurids—have an argillic horizon above the duripan
	Haplodurids—other Durids
Gypsids	Petrogypsids—petrogypsic or petrocalcic within 100 cm of the surface
	Natrigypsids—natric horizon within 100 cm of the surface
	Argigypsids—argillic horizon within 100 cm of the surface
	Calcigypsids—calcic horizon within 100 cm of the surface
	Haplogypsids—other Gypsids
Argids	Petroargids—duripan, petrocalcic, or petrogypsic horizon within 150 cm of the surface
	Natrargids—have natric horizon
	Paleargids—more than 50 cm to a contact and either an absolute clay content increase of 15% within 2.5 cm or a 7.5YR or redder argillic horizon within which the clay content does not decrease by 20% or more of its maximum within 150 cm.
	Gypsiargids—gypsic horizon within 150 cm of the surface
	Calciargids—calcic horizon within 150 cm of the surface
	Haplargids—other Argids
Calcids	Petrocalcids—petrocalcic horizon within 100 cm of the surface
	Haplocalcids—other Calcids
Cambids	Aquicambids—irrigated and have aquic conditions within 100 cm for some time in most years or saturated with water within 100 cm for 1 month most years
	Petrocambids—duripan, petrocalcic, or petrogypsic within 150 cm of the surface
	Anthracambids—have an anthropic epipedon
	Haplocambids—other Cambids

Patton are still clearly visible in the desert pavement and soils of the Sonoran Desert in southern California.

Classification of Aridisols

Aridisols are defined as having an aridic soil moisture regime, an ochric or anthropic epipedon, and one or more of the following subsurface horizons within 100 cm of the soil surface: argillic, cambic, natric, salic, gypsic, petrogypsic, calcic, petrocalcic, or duripan. These subsurface horizons, plus the cryic soil temperature regime, provide the basis for differentiation of suborders (Table 10.1; Figure 10.12).

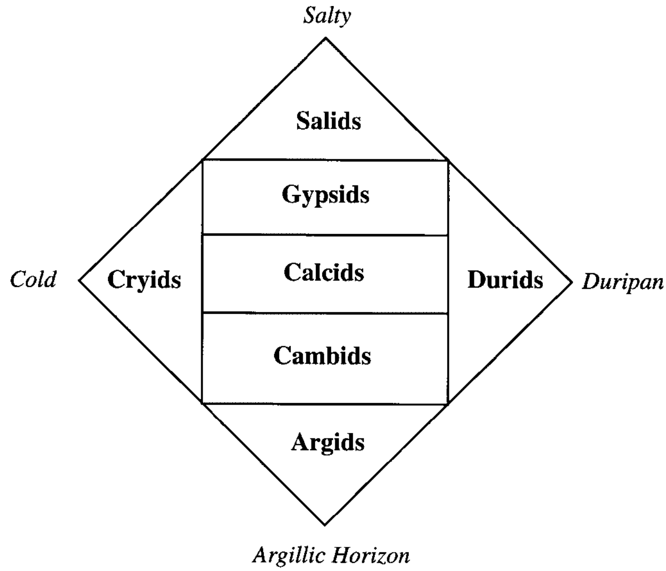

Figure 10.12. Diagram showing some relationships among suborders of Aridisols.

Prior to 1994, only two suborders existed: Argids, with an argillic or a natric horizon, and Orthids. Great groups of the Orthids were differentiated on the basis of the other diagnostic horizons now used to identify more suborders (Witty 1990).

The Cryids have a cryic soil temperature regime and occur at high latitudes and high elevations. These soils, wherein soil processes are inhibited, not only by lack of water, but also by low temperatures, are common in the intermountain region of the western United States (Hipple et al. 1990). The Argids, Durids, and Petrocalcids (petrocalcic horizon) have formed on the oldest geomorphic surfaces, as on the crests of dissected alluvial fans or on hillslopes protected from sheet erosion by vegetation or by large boulders and rock outcrops. Cambids and Calcids are found on geologically younger side slopes and surfaces of intermediate age (Figure 10.3). Gypsids and Salids often occur near playa margins, where salts are concentrated at or near the soil surface by upward flux from a water table driven by evaporation. Less commonly, they occur in association with Cambids and Calcids where parent materials are saline or where gypsum is added with eolian material (Reheis 1987; Eswaran and Zitong 1991). Soils of the youngest surfaces in dry regions, both the steep mountain slopes and recent alluvial bottoms, have not developed any diagnostic subsurface horizons and are classified as "Torri" subgroups of the Entisol order. Figure 10.4 diagrams the relationship of soils from other orders that also have aridic (torric) moisture regimes, the Torrox and Torrerts as well as the several other soils that approach the aridic moisture regime in dryness.

From a land management point of view, the presence of diagnostic horizons, especially duripans and petrocalcic and natric horizons, within 100 cm of the soil surface can pose severe limitations for land leveling and drainage for irrigation. Diagnostic horizons occurring between 100 and 150 cm are identified at the great group level.

Perspective

A century ago geographers, geologists, and pedologists presumed that formation of clay and translocation of it or anything else to the subsoil was virtually impossible in deserts because of inadequacy of precipitation. It has been a surprise, therefore, that studies of desert soils over the last half-century or so have shown that Aridisols with complex morphology, including presence of argillic horizons, duripans, and petrocalcic horizons, are common and that areas of sand dunes and rock land are not predominant. Entisols are important associates of Aridisols. The chemical and physical reactions operating in soils of humid regions also operate, although with less intensity and at shallower depths, in arid regions. Some Aridisols occur on landscapes that are more than 1 million years old, a time interval that has allowed extreme development of accumulations of clay, silica, and carbonates. These soils are polygenetic, having responded to a succession of environmental conditions that determined the extent of mineral weathering and leaching of the weathering products. Eolian additions have had a major impact on many Aridisols, by adding carbonate, clay, and salts to the soil surface, particularly where an effective dust trap in the form of a desert pavement protects the surface. Entrapped eolian material is fed into the underlying vesicular layer, thereby lifting the desert pavement and its landscape. Cycles of wetting and drying and faunal activity (termites have dismembered entire argillic horizons at some sites) have destroyed argillans in some soils. Relationships of factors of soil formation to arrangement of soil boundaries and mosaics of contrasting ecological conditions at various scales are particularly evident in arid regions. Pedoecological studies of desert landscapes may enable us to better understand soil landscape patterns of humid regions. Our awareness of the fragility of desert soils needs to be shared widely with human populations who are seeking to utilize the resources of arid lands.

Entisols: Recently Formed Soils

11

Entisols are soils that “have little or no evidence of the development of pedogenic horizons” (Soil Survey Staff 1999). Where such horizons are lacking but plant growth goes on, the plant growth is evidence enough that the unconsolidated parent materials are functioning as soils. Root domains are established in Entisols, whether or not pedogenic soil horizons have formed. Entisols are either made of such inert materials (such as quartz sand) or are of such slight and recent exposure to pedogenic processes that diagnostic horizons are typically absent, although ochric epipedons and/or albic horizons may be present. Entisols include both (1) simple soils with one or both of the horizons mentioned, and (2) pedogenically featureless, life-supporting natural materials, intermediate between nonsoil (hard rock, deep bodies of water, glaciers) and horizonated soil. Featureless soil bodies created by human mixing of soil materials are assigned to this order.

Interestingly, the full evolutionary sequence from parent material to well-developed soils is paralleled by a progressive awareness of soil by landscape painters in the Western world. The American pedologist Hans Jenny (1968) has pointed out that painters of the fourteenth century showed whole landscapes of bare rock outcrops. Artists of succeeding centuries came to portray first a vegetative cover over rock and more recently a layer of soil (horizons) between vegetation and substratum. We are concerned in this chapter with the simpler end of the soil continuum, recognizing, however, that in reality no soil is simple.

Setting

Most Entisols are on land surfaces that are very young, wet, dry, or are underlain by very resistant initial material. Entisols are commonly associated with large settlements, because notable concentrations of human population depend on productive alluvial and seashore lands and their adjacent steep uplands (which may be elaborately terraced by farmers). Also, dense populations mean disturbance of native soils by extractive activities, earth moving, and waste dumping.

Of first importance are the factors limiting soil horizon development in wetlands, alluvial lands, sandy lands, higher-lying rocky lands, and various unconsolidated deposits such as loess and mudflows. The global distribution of Entisols

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

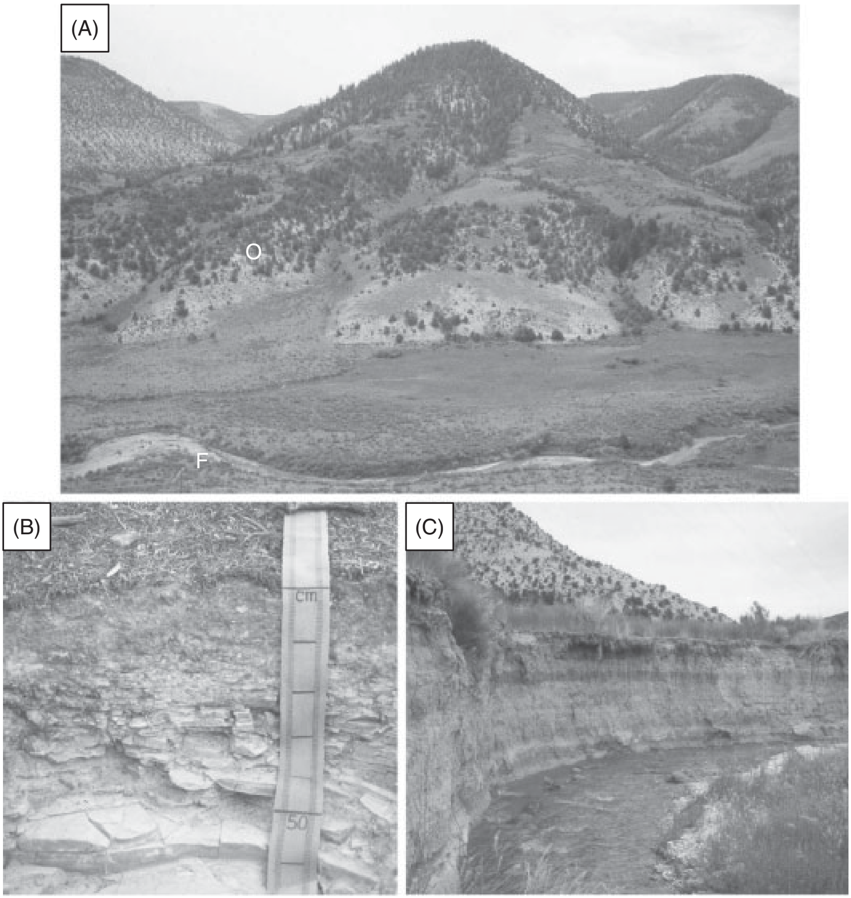

Figure 11.1. (A) Landscape view in the mountains of eastern Utah. (B) Orthents occur under pinyon and juniper trees on south-facing slopes as indicated by "O" in the landscape photograph. Note shallow depth to shale bedrock. (C) Fluvents occur in the valley bottom, as indicated by "F" in the landscape photograph. Note the buried A horizons that give rise to an irregular decrease in organic carbon content with depth. Mollisols are found on low-gradient uplands and north-facing slopes.

indicates that many factors are involved, including the following, which may operate in various combinations:

1. Arid macroclimates and microclimates, either cold or warm, may limit the amount and duration of water movement in the soil and biotic influence on the soil.
2. Mass wasting and other forms of erosion may remove material from a site as fast as or faster than most pedogenic horizons form. Steep slopes favor the operation of this process. In the tectonically active San Gabriel Mountains of southern California, Entisols predominate on very steep (45–85%), xeric, chaparral-covered slopes. These soils form in a thin, quasi-stable colluvium mantle that is no more than 400 years old (Graham et al. 1988). A combination

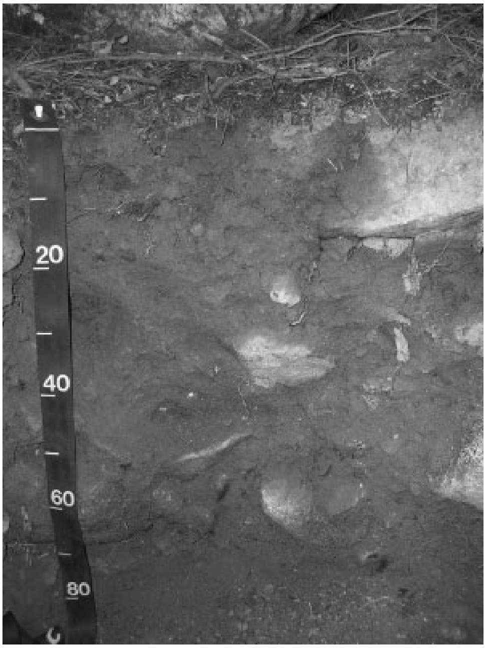

Figure 11.2. Photograph of a sandy-skeletal, mixed, mesic Typic Xerorthent formed in a 75-year-old debris flow deposit in the San Bernardino Mountains, California (Turk et al. 2008). Note the O horizon formed by the accumulation of organic matter on the surface, a weak A horizon in the upper 10 cm, and lack of other distinct horizons. (Photo by J.K. Turk) For color detail, please see color plate section.

of aridity and slope appear responsible for many Entisols, such as the Orthents on steep south-facing slopes in eastern Utah depicted in Figure 11.1A and B.

3. Cumulization may add new material to the surface of the soil as fast as or faster than the new material can be assimilated into a pedogenic horizon. This process is at work on alluvial floodplains (Riecken and Poetsch 1960; Daniels 2003), debris and mud-flow deposits (Dickson and Crocker 1954; Turk et al. 2008), deltas, footslopes, areas of active loess and dune sand deposition, and shore lands bordering lagoons and estuaries. Removal of vegetation, whether by natural or anthropic means, from part of a landscape accelerates erosion and related depositions.
4. Paucity of soil plasma in inert materials such as highly siliceous sediments, or immobilization of soil plasma in carbonate-rich flocculated materials, inhibits profile differentiation by illuviation.
5. Exceptional resistance to weathering (pedologic inertia) of some initial materials, such as quartzite bedrock, prolongs the period of undistinguished horizonation.
6. Infertility and toxicity of some initial materials to plant growth limits biogenetic differentiation of the soil profile. In serpentine barrens, Entisols may be associated with bedrock outcrops and lithic subgroups of other soil orders.
7. Saturation with water or even submergence of the soil for long periods inhibits the kind of horizon development found in soils of other orders. Nevertheless, distinctive soil development does occur in subaqueous soils (Demas and Rabenhorst 1999), and it varies by submerged geomorphic settings (Bradley and Stolt 2003).

8. Shortness of time since exposure of initial materials to the active factors of soil formation limits profile development (Figure 11.2). Fresh lava flows, volcanic ash deposits, marine or lacustrine flats newly exposed by uplift of land or by lake drainage, and avalanche and landslip scars provide sites for very young soils, as do situations mentioned in item 3 above. Human manipulation of soil and geologic materials with bulldozers and other machines leaves fresh initial materials exposed to active pedogenic factors (Smith 1976).
9. A recent drastic change in the biotic factor may initiate formation of a different soil profile in an old soil profile, which serves as initial material. For example, human activities may cause hemlock forests to be replaced by aspen and bracken. This may allow degradation of the spodic horizon to the point that the soil no longer qualifies as a Spodosol and becomes an Entisol, all within a period of less than a century (Hole 1976).
10. Extensive burrowing and digging by animals mixes soil material and accelerates erosional losses on slopes. This bioturbation can retard diagnostic horizon formation and maintain soils as Entisols (Weitkamp et al. 1996).

Topolithosequences of Entisols may occur in floodplains of rivers, ranging from sandy soils (Psamments) of islands, bars, and riverbanks to Alluvial soils of finer texture (Fluvents) on the natural levees and wet silty soils (Aquents) in the backwater lowlands. Of the 590 million hectares of Alluvial soils (Alluvial as defined in the 1938 classification [Baldwin et al. 1938]) in the world (Kellogg and Orvedal 1968) that support about one-third of the human population, perhaps less than half are Entisols, the remainder being Vertisols, Mollisols, and Inceptisols. Entisols cover 8.6% of the land area in alluvial, mountainous, desert, and sandy regions. Because Entisols are associated in many areas where soils of other orders predominate, Entisols may actually occupy 20% of the land area of the planet.

Pedogenic Processes

The emphasis of the previous section has not been on the nature of the processes, but on the curtailment of their operation in Entisols. This can amount to a complete absence of some processes at certain sites. For example, oxidation may be excluded from saturated soils; cryoturbation from climatic zones in which there is no freezing; cutan formation from sandy soils lacking plasma.

The full range of soil formation processes listed in Table 5.1 commonly operates to some degree in Entisols of the world; however, the impact of these processes has not been great enough to produce soil features recognized as diagnostic for other orders. For example, melanization is usually active to a limited degree in all Entisols as soon as plants grow and add carbon from their decaying tissues to the mineral soil material. Melanization contributes to the formation of the ochric epipedon but has not contributed enough carbon to form mollic, histic, or umbric epipedons that are excluded by definition from the Entisol order. Examples of this phenomenon are found in the lysimeter soils at the San Dimas Experimental Forest in southern California. After 41 years of soil formation under monocultures of native shrubs, the incorporation of organic matter darkened A horizons, but they were not sufficiently thickened to be mollic epipedons, so the soils remained Entisols (Graham and Wood 1991). Other examples of this incipient melanization are seen in a debris flow chronosequence in the San Bernardino Mountains, California (Turk et al. 2008). Thin A horizons form within decades (Figure 11.2), but most of the organic matter contributed by the coniferous forest accumulates as O horizons on the soil surface so that even after 244 years, the soils still have ochric epipedons and little other morphologic development.

Fragmentation of diagnostic horizons by human manipulation of soils can create profiles with no diagnostic horizons. This is a basis for definition of the suborder Arents. Buried horizons, such as ochric epipedons, are observed in profiles of many cumulative Entisols (Fluvents). Some Lithic Entisols (Orthents, in particular) may be unstable soils that are prevented by erosion from developing profiles that have horizons that are more distinct.

Grossman (1983) has pointed out that change in initial compactness and the effect of this change on pedogenic processes are important to understanding Entisols. Sediments that accumulated under water have a low compaction state that can be given numerical ratings (liquid index of engineers; *n*-value of Dutch workers). High porosity (low compaction) is markedly and, to some extent, irreversibly reduced by drainage, natural or artificial. Capillary forces reorient clay particles in the drying process. Because the number of wet–dry cycles increases with proximity to the ground surface, bulk density of Entisols increases upward at some sites. At other sites the opposite trend may prevail because of effective bioturbation, wetting-induced intra- and inter-particle swelling that disrupts rock fabric, or because of relaxation of the sediment in response to removal of overburden from compact glacial tills, mudflows, or other material. Traffic of machinery, animals, and people produces high states of compaction in Entisols, as they do in soils of any order.

The ubiquity of textural discontinuities in Entisols formed in floodplains means that water movement is obstructed, even when coarse material underlies fine material (Miller 1973). In many Entisols, horizontal planar voids predominate over vertical ones. Bouma et al. (1977), using dyes as they flooded clayey soil cores, found evidence of less water movement in a Fluvaquent than in an Inceptisol.

The usual irregularity of organic carbon distribution with depth in Fluvents did not occur (prior to construction of an Aswan Dam) in the Nile River valley and delta, where the soil nitrogen curve paralleled those of prairie soils (Figure 7.16 of Jenny 1980). The reason for this smooth curve is that annual increments of 1 mm of alluvium had a constant content (0.1–0.2%) of nitrogen, and decomposition of organic matter was progressive in buried lamellae.

Shallow Entisols in pinyon-juniper woodlands on the Colorado Plateau do not show regular decreases in organic carbon with depth because the decay of tree roots concentrated above lithic contacts is a major source of organic matter (McDaniel and

Graham 1992). Under tree canopies, litter contributes organic matter to the A horizon, so organic carbon content is highest near the surface and where the roots concentrate just above the lithic contact. Between trees, the soil surface is barren and eroded, but the roots of the trees extend laterally and proliferate above the bedrock, so organic carbon content is entirely a product of roots and increases with depth in these soils.

Entisols that contain an abundance of rock fragments may show a decrease upward in the volume of this coarse fraction, even in the uppermost 25 cm of soil. There may be a concomitant increase in porosity of the surviving cobbles, which may actually store and supply plant-available water, even though plant roots cannot penetrate them (Schafer et al. 1980). Porous, weathered rock fragments in soils support active microbial populations, and plants can obtain both water and nutrients from them (Agneli 2001; Ugolini et al. 2001). Abundance of rock fragments is reported to reduce the erodibility of Entisols, the same as in stony soils of other orders.

Subaqueous Entisols (Wassents) occur in estuarine sediments under as much as 2.5 m of water. They are subject to distinct pedogenic processes involving addition of shells and organic matter, losses of organic matter and surficial material, transfer of oxygen through diffusion and bioturbation, and transformation of organic matter and sulfur (Demas and Rabenhorst 1999).

Uses of Entisols

Entisols present engineering problems in many regions. Erosion by water, wind, and mass wasting is important in steep and hilly to mountainous areas, where runoff is rapid. Rocky, shaley, and sandy lands pose these hazards in different ways. Flooding and deposition must be reckoned with in lowlands, particularly on river floodplains and tidal flats. The boundaries of areas endangered by flooding are usually the boundaries of certain alluvial and shore land soil bodies, Entisols included. Drainage of wetlands involves notable areas of Aquents, including some of the “cat clays” (Sulfaquents of tidal marshes), which become extremely acid upon drainage through oxidation of sulfides (Pons 1969; Vlek 1971; Fanning and Fanning 1989).

Inhospitable lands, including terrains of Entisols and associated salt flats and rock outcrops, may be assigned a value as wildlife preserves or at least as wilderness areas with aesthetic qualities. The Orthent-Fluvent–Mollisol soilscape shown in Figure 11.1 is grazed by cattle, deer, and elk in the uplands and antelope in valley bottoms. Minor areas of Fluvents on this eastern Utah landscape are cultivated for hay. Military testing operations by various nations of the world have been concentrated to a notable extent on Entisols. Cattle and sheep range in subhumid, semiarid, and arid regions that cover vast areas of soils of this order, including parts of the Sand Hills of Nebraska.

Millions of people support themselves on fertile and workable Entisols developed from alluvium (of the Ganges Plain and Mekong Delta, for example) and shore deposits in association, in many places, with soils of other orders. Entisols formed by active cumulization of fertile soil material are highly prized in primitive agriculture because of their ability to grow crops without fertilization. Flooding often limits agricultural practices on these soils (McCormack 1971). Changes of water table levels resulting from flood management structures may affect soil productivity drastically and may even affect the classification of the soils. General farming, with and without irrigation, and specialized farming, such as for alfalfa production, are practiced on level, fertile Entisols. With irrigation, Torrifluvents and Torripsamments in the San Joaquin, Coachella, and Imperial valleys of California annually produce crops worth billions of dollars, including citrus, grapes, melons, cotton, hay, nursery stock, turf, and a wide variety of vegetables. Citrus groves are on Psamments in Florida, and some truck farms are on lithic silty "marl land" Orthents. Rice paddies and other forms of cultivation requiring control of surface water are facilitated by the level terrain associated with the water-deposited Entisols (Fluvents). Quartzipsamments have an extremely low content of plant nutrients except where these are imported, as they are by salt spray in the "sunken forest" on a barrier island off Long Island, New York, where precipitation is adequate to prevent accumulation of sodium (Art et al. 1974). In other Quartzipsamments residual, nonexchangeable K is slowly released from hydroxy-interlayered vermiculite as it is weathered by low concentrations of oxalate in these acid soils (Comerford et al. 1990).

Psamments and other coarse-textured Entisols are particularly susceptible to leaching of pesticides and other environmental contaminants. Soil survey data can be used to effectively identify these sensitive soils so that special management practices can be used to protect groundwater (Teso et al. 1988).

The presence of unpredictable lithologic discontinuities in Fluvents, created by irregular flood intensities, requires careful observation of subsoil during mapping and makes interpretation of mapping units difficult.

Classification of Entisols

Entisols are soils without properties that are diagnostic of the other orders. They support or are capable of supporting plants and are differentiated from non-soil materials such as bare rock, deep water, or ice at the surface of the earth. Besides ochric and anthropic epipedons, albic and agric horizons, Entisols may have fragments of diagnostic horizons that are not arranged in any discernable order. Some Entisols may have thin cambic horizons at a shallow depth (above 25 cm).

Figure 11.3 arranges the five suborders to show relationships to some factors and properties. These suborders may be defined briefly as follows:

1. Wassents have a positive water potential at the soil surface for more than 21 hours per day throughout the year; that is to say, they are subaqueous soils.
2. Aquents are other permanently or seasonally wet (saturated) Entisols. Even if artificially drained, they display bluish gray redoximorphic features. Sulfaquents in tidal marshes of Maryland are tentatively identified in the field by the intensity of H_2S odor (Darmody et al. 1977).

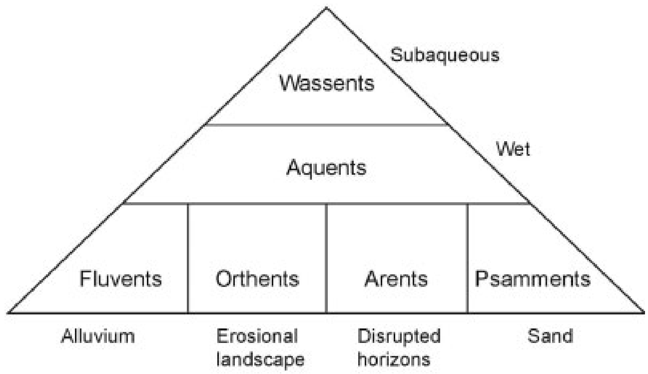

Figure 11.3. Diagram showing some relationships among suborders of Entisols.

3. Arents are Entisols that are better drained than Aquents (lacking their redoximorphic features) and that exhibit fragments of diagnostic horizons below the Ap horizon. The disturbance of the deeper horizons has often been by deep plowing one or more times, but not often enough to produce a homogenized deep Ap horizon.
4. Psamments have textures of loamy fine sand or coarser and are better drained than Aquents.
5. Fluvents are loamy and clayey (finer in texture than loamy fine sand) alluvial soils with very simple profiles. Irregularity of organic matter content with depth is diagnostic. Stratification is common in alluvium and in soils derived from it (Figure 11.1C).
6. Orthents are loamy and clayey Entisols (better drained than Aquents), with a regular decrease in content of organic matter with depth (Figure 11.1B, Figure 11.2). They are commonly on recent erosional surfaces, but they also occur on recent thick eolian, glacial, or mass-wasting deposits (Soil Survey Staff 1999).

The suborders are subdivided into great groups (Table 11.1) on the basis of several factors: mean annual soil temperature and range of soil temperature, sandiness and quartz content, stratification, presence of sulfidic material, and low-bearing capacity (Soil Survey Staff 1999).

There have been proposals for classifying soils created by intense human activity within the Entisol order. For soils where all fragments of preexisting diagnostic horizons have been destroyed and human-manufactured objects are common in the pedons, Urbic, Spolic, Garbic, and Scalpic subgroups have been suggested (Fanning and Fanning 1989). Short et al. (1986), while classifying and mapping soils of the Mall in Washington, D.C., found 74% of the area occupied by such Entisols, the remainder being classified as Inceptisols. A like proportion of intensely disturbed Entisols (Udorthents) was mapped in Central Park in New York City (Weber et al. 1982), and in minelands of Pennsylvania (Ciolkosz et al. 1985). By far the most interest in such soils has been in surface-mined land, where careful classification of the

Table 11.1. Suborders and great groups in the Entisol order

Suborder	Great Groups
Wassents	Frasiwassents have very low salinity throughout upper 100 cm.
	Psammowassents have sand and loamy sand to the 100 cm depth.
	Sulfiwassents have ≥15 cm thick zone of sulfidic materials within 50 cm of surface.
	Hydrowassents—very soft; low-bearing capacity (high *n* value).
	Fluviwassents—organic carbon ≥0.2% at 125 cm depth and/or an irregular decrease in organic carbon between 25 and 125 cm.
	Haplowassents—other Wassents.
Aquents	Sulfaquents have sulfidic materials within 50 cm of the surface.
	Hydraquents—very soft; low-bearing capacity (high *n* value).
	Gelaquents have a gelic soil tempreature regime.
	Cryaquents have a cryic soil temperature regime.
	Psammaquents have sand and loamy sand to the 100 cm depth.
	Fluvaquents—slope <25%, organic carbon ≥0.2% at 125 cm depth and/or an irregular decrease in organic carbon between 25 and 125 cm.
	Epiaquents have episaturation.
	Endoaquents—other Aquents (have endosaturation).
Arents	Ustarents have an ustic soil moisture regime.
	Xerarents have a xeric soil moisture regime.
	Torriarents have an aridic soil moisture regime.
	Udarents—other Arents (have a udic soil moisture regime).
Psamments	Cryopsamments have a cryic soil temperature regime.
	Torripsamments have a torric (aridic) soil moisture regime.
	Quartzipsamments—sand fraction is >90% resistant minerals.
	Ustipsamments have an ustic soil moisture regime.
	Udipsamments—other Psamments (have a udic soil moisture regime).
Fluvents	Gelifluvents have a gelic soil temperature regime.
	Cryofluvents have a cryic soil temperature regime.
	Xerofluvents have a xeric soil moisture regime.
	Ustifluvents have an ustic soil moisture regime.
	Torrifluvents have a torric (aridic) soil moisture regime.
	Udifluvents other Fluvents (have a udic soil moisture regime).
Orthents	Gelorthents have a gelic soil temperature regime.
	Cryorthents have a cryic soil temperature regime.
	Torriorthents have a torric (aridic) soil moisture regime.
	Xerorthents have a xeric soil moisture regime.
	Ustorthents have an ustic soil moisture regime.
	Udorthents—other Orthents (have a udic soil moisture regime).

new soils is needed for land-use planning (Thurman and Sencindiver 1986). Workers have reported special pedogenic features such as the "fritted structure" (semifused or semiintegrated aggregates loosely compressed together) formed by action of mining wheels and belt transportation of mine spoils (McSweeney and Jansen 1984). Areas of such soil are, however, relatively small and lack homogeneity, and taxonomic recognition has not been established.

Perspective

Entisols are those soil materials that support (or are capable of supporting) plant growth, but which show little or no pedogenic horizon development. These soils are important components of soilscapes used by human populations whose activities daily create more such soils. Processes of erosion and sedimentation also continually create new Entisols. Genetic changes are rapid in the many Entisols that are quite porous and are in humid and subhumid climatic zones. This means that if their surfaces are protected from erosion or deposition, a certain number of hectares of these soils may “graduate” each century into another soil order. Entisols that are in equilibrium with the environment, forming as rapidly as erosion depletes them (Whittaker et al. 1968), constitute one kind of “climax” soil community.

The study of Entisols challenges pedologists to observe weakly developed soils carefully and to consider their genesis by geologic and anthropic processes, and the initial impacts of biological influences. The development of pedogenic features is usually subtle, but close attention to details will reveal the nature of influential soil processes.

Gelisols: Very Cold Soils

12

Gelisols (from the Greek *gelid*, very cold) are soils that contain permafrost within 200 cm of the ground surface. Permafrost is a soil climatic condition under which organic and inorganic soil materials, ice and rock, continuously have temperatures at or below 0°C. Muller (1947) proposed the term *permafrost* for this condition that has persisted in polar and alpine regions for centuries to millions of years. The Gelisols are therefore characterized not by a diagnostic soil horizon, but rather by a diagnostic perennial coldness. Permafrost-affected soil horizons in Gelisols encompass a wide range in characteristics; they range from very dry and loose in arid regions to ice-cemented in areas that have effective precipitation. Gelisols were officially recognized as an order in *Soil Taxonomy* with the 8th edition of *Keys to Soil Taxonomy* (Soil Survey Staff 1998). Previously, *Soil Taxonomy* had identified these soils as pergelic subgroups of Entisols, Inceptisols, Histosols, Mollisols, and Spodosols (Rieger et al. 1979) or Gelicryands (Soil Survey Staff 1994). Cryosols, one of the 32 reference soil groups of the *World Reference Base for Soil Resources* (IUSS 2008), are essentially equivalent to Gelisols.

Setting

The one prerequisite for Gelisols is the presence of permafrost within 2 meters of the ground surface. Permafrost is defined by *Soil Taxonomy* as a thermal condition in which soil material remains below 0°C for at least two or more consecutive years (Soil Survey Staff 2010a). This condition is fulfilled in the two polar and circumpolar regions and at high elevations in mountains at lower latitudes. Figure 12.1 (Péwé 1991) and Figure 12.2 (Bockheim 1995) show the distribution of areas of continuous (occupying ≥90% of an area) and discontinuous (≤90% of an area) permafrost.

The global extent of Gelisols is not as well documented as that of other soil orders. Estimates range from ~11.8 million km^2 or 9% of the ice-free land area (USDA-NRCS database) to 18 million km^2 or ~13% of the ice-free land area (Bockheim 2002). In terms of millions of km^2 of permafrost-affected soils, Russia has about 11.7, Canada 4.0, the United States 1.8 (80% of Alaska has permafrost), and Antarctica 0.5. Large areas of Gelisols are also found in China and Mongolia; smaller areas occur in northern Europe and Greenland (IUSS Working Group WRB 2006).

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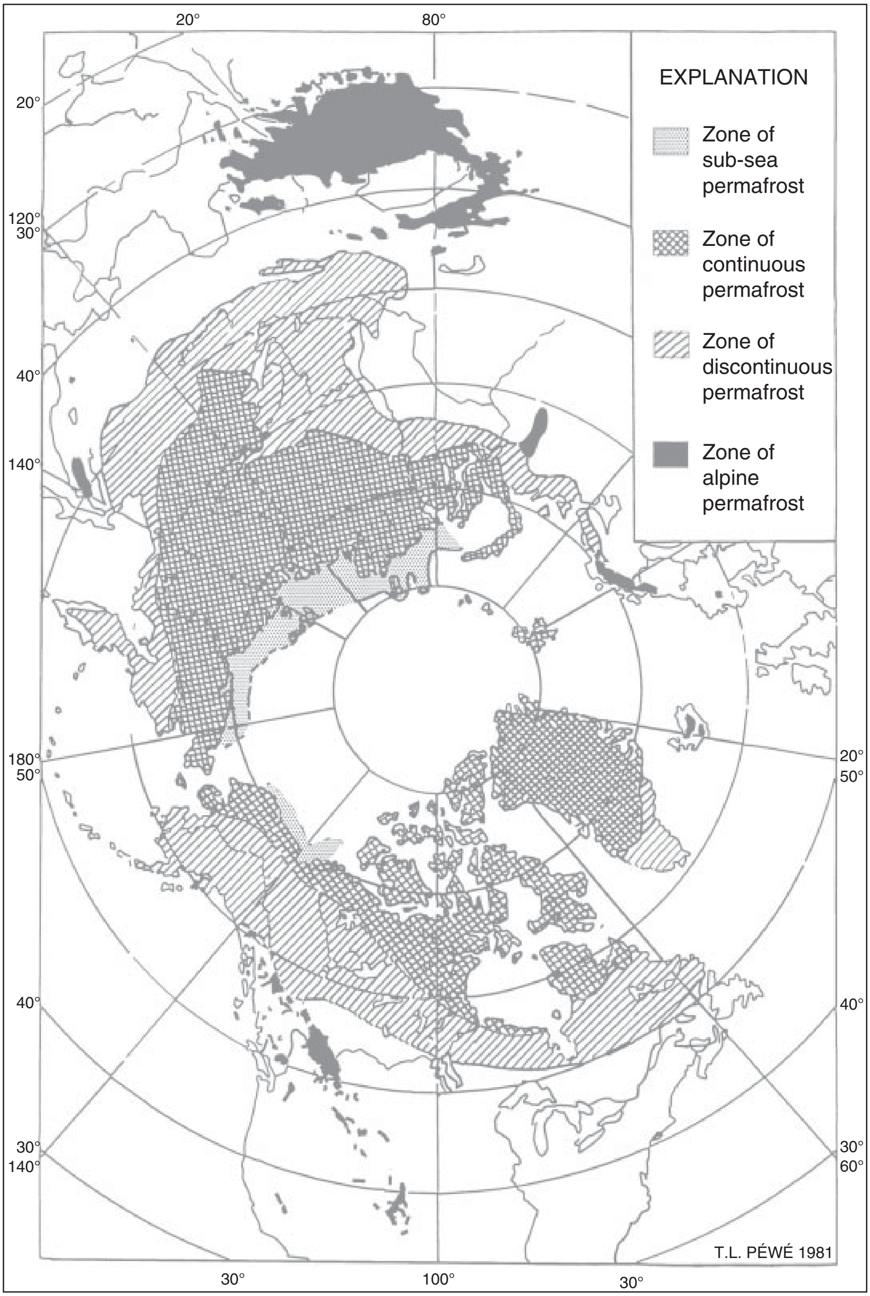

Figure 12.1. Distribution of permafrost in the Northern Hemisphere. Isolated areas of alpine permafrost not shown on the map may exist in the high mountains outside the map area. It is probable that only a small part of the black alpine areas of the map is actually occupied by permafrost. (Map after Pewe 1991, reproduced with permission)

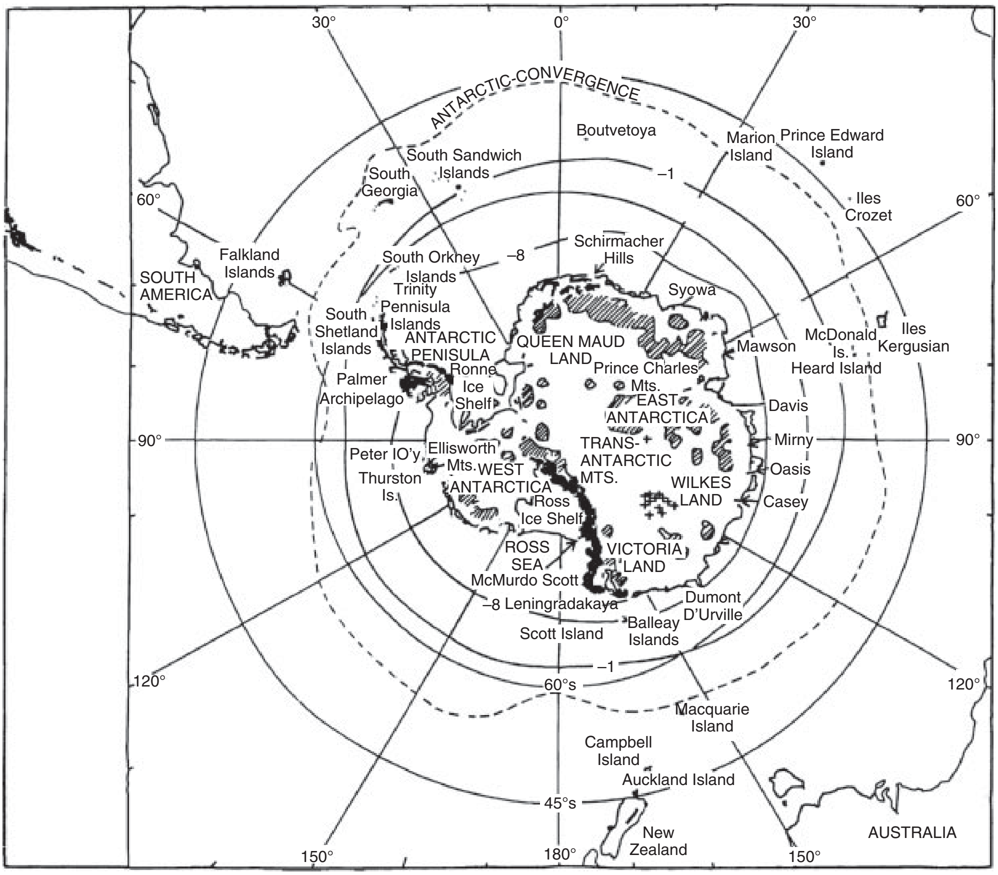

Figure 12.2. Distribution of permafrost in the Southern Circumpolar Region (~>60°S latitude). The black areas along the Antarctic Peninsula and the Trans-Antarctic Mountains are ice free. The −8°C mean annual air temperature isotherm is the approximate northern limit of presumed continuous permafrost in soils. The striped areas have subglacial permafrost. The blank (white) areas of the continent are subglacial permafrost-free because of the enormous pressure of the thick glacial ice. The crosses mark the known sites of subglacial meltwater lakes. The Antarctic Convergence is the zone of mixing of the circumpolar cold waters and the subtropical warm waters, which brings up nutrients. In this zone, whales, fish, and crustaceans feed. (Map by J.G. Bockheim 1995, reproduced with permission)

Patterned ground is common in terrain featuring permafrost, and repetitive arrangements of materials include unsorted and sorted circles, nets, polygons, steps, and stripes (Figures 12.3, 12.4). Frost cracking of the land surface into polygons results from shrinkage of the ground during very cold winters. Water from the active layer in summer seeps into cracks and freezes, starting the growth of vertical ice wedges, which may take centuries to reach a width of a meter or so at the top of the permafrost that is at the base of the active layer. Within the miniature terrains of units of patterned ground, the force of gravity moves stones from frost-heaved ridges

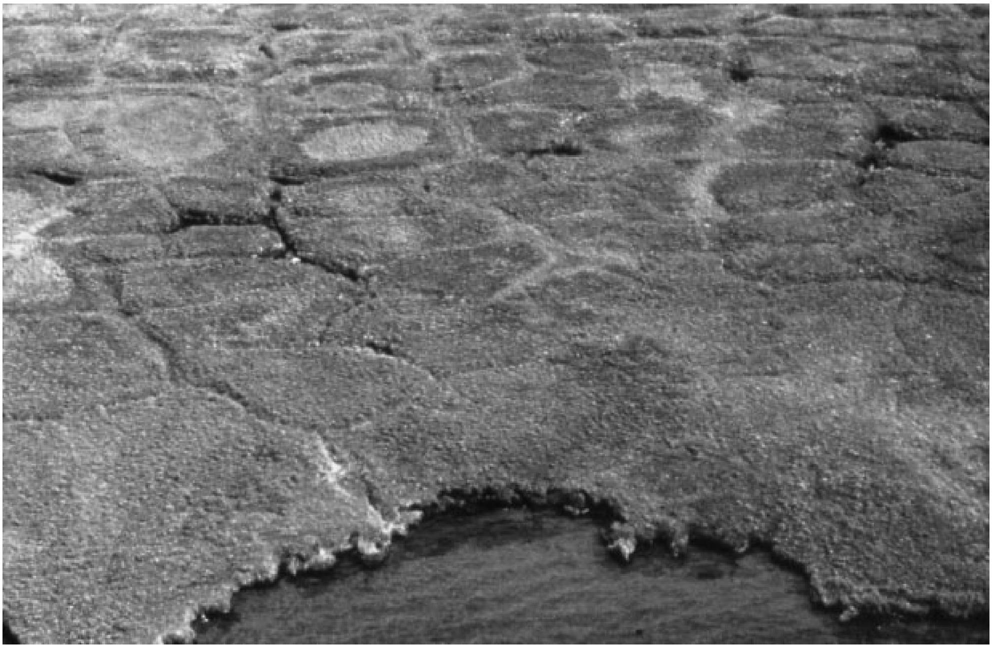

Figure 12.3. Gelisol landscape near Prudhoe Bay, Alaska showing patterned ground. Polygons are approximately 10 m across. (Image courtesy of Dr. Maynard Fosberg, University of Idaho.) For color detail, please see color plate section.

to collapsed depressions. Workers also report frost boils and mounds, palsas (small pingos), pingos, peat rings, and beaded drainage patterns (Tedrow 1977; Campbell and Claridge 1987; Gavrilova 1993).

In the Northern Hemisphere, Gelisols are largely associated with tundra and areas of boreal forest or taiga. Typical tundra vegetation consists of dwarf trees and shrubs, grasses, sedges, lichens, and mosses. The tundra vegetation provides a food source for herds of reindeer (genus *Rangifer*, called caribou in North America) that subsist on lichens in their migrations and small herds of Musk Ox (*Ovibos maschatus*) that feed on grass from 64° N to shores of the Arctic Ocean. Annual precipitation is often relatively low; the arctic regions of Alaska that are dominated by Gelisols receive an average of 125 to 255 mm annual precipitation (USDA–Natural Resources Conservation Service 2006).

The inland dry valleys of Antarctica provide a much different setting. Annual water-equivalent precipitation ranges from less than 10 mm to as much as 100 mm (Doran et al. 2002). Here, plants are absent and the soil cover is desert pavement, which is a natural layer of ventifacted and "varnished" pebbles that protect the soil from wind erosion. One such surface in Antarctica has scattered dark basaltic pebbles that were erupted from a volcanic cone 2.7 million years ago (Campbell and Claridge 1987), providing evidence for very long-term landscape stability. The

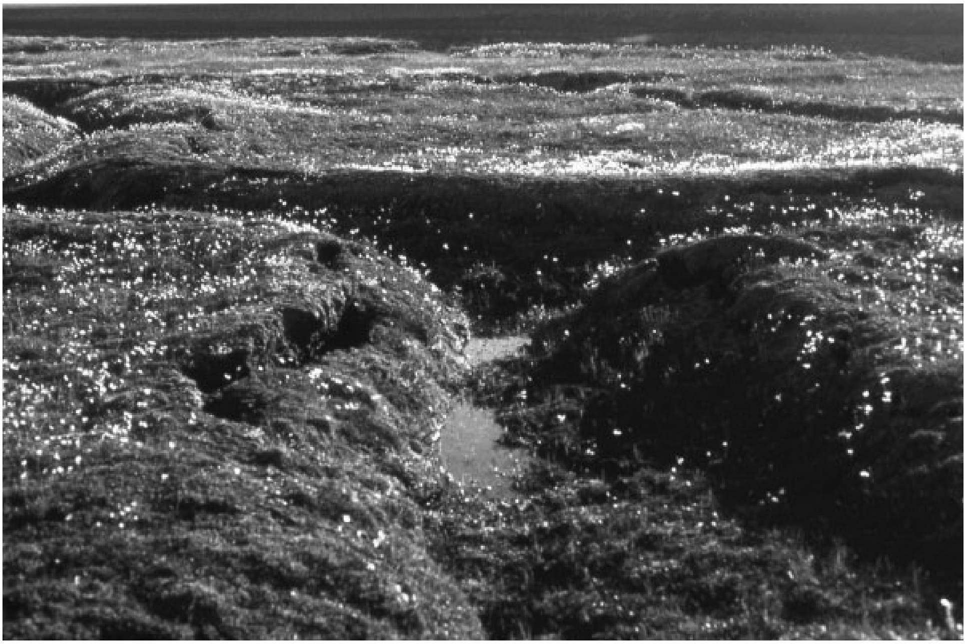

Figure 12.4. Close-up view of polygons shown in Figure 12.3 and are delineated by troughs up to 1 m deep. (Image courtesy of Dr. Maynard Fosberg, University of Idaho.) For color detail, please see color plate section.

moister coastal areas support cryptogamic vegetation, consisting primarily of mosses and lichens (Cannone et al. 2008).

Pedogenic Processes

In the very cold environments where Gelisols are found, pedogenic processes such as weathering, gains and losses, and transformations and translocations of mineral and organic materials proceed slowly and are often very weakly expressed. Some of the Gelisols from the dry valleys of Antarctica have challenged our traditional concepts of soil; these unconsolidated materials do not support higher plant life, nor do they exhibit clear morphological evidence of weathering and organic matter accumulation. However, several decades of research have demonstrated that soil-forming processes—redistribution of soluble salts, reddening due to chemical weathering, formation of desert pavement—have been operative in these soils (Ugolini and Bockheim 2008). Data from an example of such a Gelisol (Calcic Anhyorthel) are provided in Table 12.1. Little horizon differentiation is seen, clay contents are quite low, and weathering, as indicated by secondary Fe and reddening, is weakly expressed.

Gelisols are characterized by an active layer that overlies permafrost. During summer months, the upper few centimeters or several decimeters of a pedon are thawed, and the active layer extends to its greatest depth. At the bottom of the

Table 12.1. Selected morphological, physical, and chemical properties of Arctic and Antarctic Gelisols. Data for the Calcic Anhyorthel (Pedon. no. 99P0325) and the Terric Hemistel (Pedon no. 99P0537) are from the USDA-NRCS National Cooperative Soil Survey Soil Characterization Database (Soil Survey Staff 2010b)

Horizon	Depth cm	Moist color	Textural class	Clay %	pH H_2O	Total Carbon %	CEC pH7 cmol (+) kg^{-1}	Secondary Fe[a] %	Electrical conductivity dS m^{-1}
Calcic Anhyorthel—Antarctica									
A	0–3	10YR 3/2	coarse sand	1.6	8.1	0.9[b]	6.5	0.3	67.8
C1	3–15	10YR 3/3	sand	0.8	8.5	0.9[b]	8.1	0.5	12.2
C2	15–32	10YR 3/3	loamy coarse sand	1.3	8.0	0.9[b]	7.3	0.6	13.2
C3	32–45	10YR 3/3	loamy coarse sand	2.6	7.8	1.5[b]	6.4	0.4	12.1
C4	45–69	10YR 4/3	loamy coarse sand	3.7	7.8	2.2[b]	7.3	0.4	11.1
Terric Hemistel—Alaska									
Oe	0–40	5R 2/2	–	–	6.6	49.3	184.8	0.4	–
Bw	40–60	10YR 3/3	clay loam	29.6	6.9	4.8	29.1	2.0	–
Cf	60–90	10YR 2/1	loam	21.5	7.3	4.8	20.4	1.2	–

[a]estimated using dithionite-citrate extract.
[b]includes carbon present as $CaCO_3$.

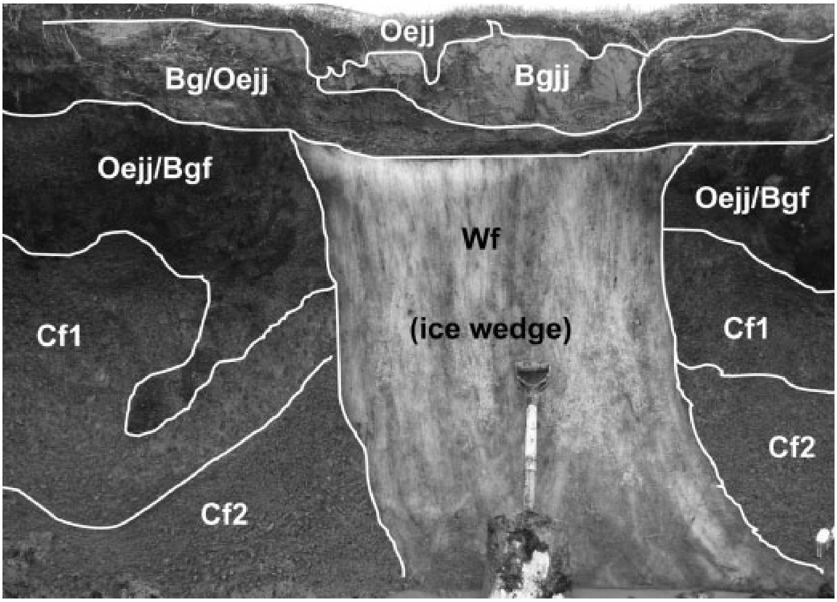

Figure 12.5. Gelisol profile from Beaufort Sea coast of northern Alaska. Soil is formed in glaciomarine deposits; vegetation consists primarily of sedges and mosses. (Image courtesy of Dr. C-L Ping, University of Alaska)

active layer is the upper surface of the permafrost, called the permafrost table. The warm-season active layer overlying permafrost ranges in thickness from 50 to 150 cm in the Antarctic islands and maritime East Antarctica and between 15 and 50 cm in interior Antarctica. Refreezing of the moist soil may be by simultaneous, slow upward extension of cementing ice above the permafrost table and downward extension of surface freezing ground. Subsoil between these two approaching freezing fronts develops a massive condition from centuries of this seasonal compaction.

One of the dominant soil-forming processes in Gelisols is cryoturbation, which is the frost mixing of soil. Many Gelisols in Arctic areas receive sufficient moisture to facilitate cryoturbation, while comparatively little occurs in drier Gelisols of Antarctica. Cryoturbation disrupts horizons, incorporates organic matter into lower horizons, and orients stones. Freezing and thawing form granular, platy, and vesicular structures in surface mineral horizons, and blocky, prismatic, and massive structures in subsoil. Ice lenses may form close to the permafrost table. Figure 12.5 is an example of a permafrost-affected Gelisol from Alaska. In this profile, the active layer extends to a depth of approximately 60 cm at the base of the Bg/Oejj horizon, and the permafrost table corresponds to the top of the ice wedge (Wf). Evidence of cryoturbation is indicated by the *jj* horizon subordinate distinction and can be seen as the irregular and broken horizon boundaries in the profile.

Gelisols support a variety of life forms that provide organic inputs. Plants are conspicuously absent from many of the driest and coldest Gelisols of Antarctica. In some such cases, the very small amounts of organic detritus present in soils likely originated from cyanobacterial activity (Hopkins et al. 2008). With relatively higher temperatures and precipitation, Gelisols have accumulations of organic matter supplied by plants (lichens, mosses, liverwort, sedges, grasses) and various animals. Soil organisms present include bacteria (N-fixing *Azotobacter,* among them), fungi, actinomycetes, mites, collembola, nematodes, tardigrades, protozoa, and algae. Gold and Bliss (1995) describe a biological surface crust on Devon Island, N.W.T., Canada consisting of cyanobacteria, fungi, lichens, and mosses that forms a protective mat, providing almost 100% soil cover.

Ornithogenic inputs can be important in some Gelisols. Additions of excreta from various seabirds strongly influences soil organic matter, soil fertility, and chemical weathering processes. In Antarctica, penguins (*Aptenodytes* sp.; *Pygosceles* sp.) breed in winter in vast rookeries. Numerous studies have documented more soil organic matter, higher contents of K, Ca, Mg, and lower pH in soils associated with these rookeries compared to adjacent unaffected areas (Michel et al. 2006). Leachates from the guano also enhance weathering of underlying rock.

Organic matter accumulation in Gelisols varies considerably according to their environmental setting. Very little organic carbon is present in soils under very cold and dry conditions such as those found in parts of Antarctica. Aislabie et al. (2008) report organic carbon contents ranging from 0.02–0.06% in surface layers of dry Gelisols from Wright Valley, Antarctica. At the other end of the spectrum, many Gelisols have organic horizons containing in excess of 12–18% organic matter by weight. An example of such a soil is a Terric Hemistel from Alaska, the data for which are presented in Table 12.1; the Oe horizon contains 49.3% total carbon on a weight basis.

Organic decomposition is very slow in Gelisols. Debris from Scott's 1910–1913 expedition to Antarctica, including horse manure and straw, is still on the ground near his hut. Organic matter decomposition is inhibited by a combination of factors including low temperatures, short thaw season, saturated conditions above the permafrost layer, and plant material that is resistant to decomposition (Hofle et al. 1998).

Globally, large quantities of carbon are sequestered in Gelisols. Data from Schlesinger (1997) suggest that despite low rates of carbon accumulation in soils of tundra ecosystems, mean carbon contents of these soils are surpassed only by those of swamp and marsh ecosystems. It is estimated that soils with permafrost contain approximately 25% of the global soil organic carbon pool (Lal and Kimble 2000). Permafrost underlying the active layer of Gelisols may contain considerable additional carbon as well. Approximately 53% of the soil organic C in Gelisols of northern Alaska is contained in the near-surface permafrost (to a 1-m depth) (Bockheim et al. 1999). Furthermore, 36% of the total carbon pool in soils of this region occurs between depths of 100 and 200 cm (Bockheim and Hinkel 2007). With warming of arctic regions and increased melting of permafrost, these forms of organic carbon could be subject to oxidation, thereby resulting in production of carbon-based greenhouse gases.

Accumulation of salt is a dominant process in many Gelisols. Marine aerosols are the chief source of salts, but salt grains (gypsum, calcite, halite, etc.) are also carried

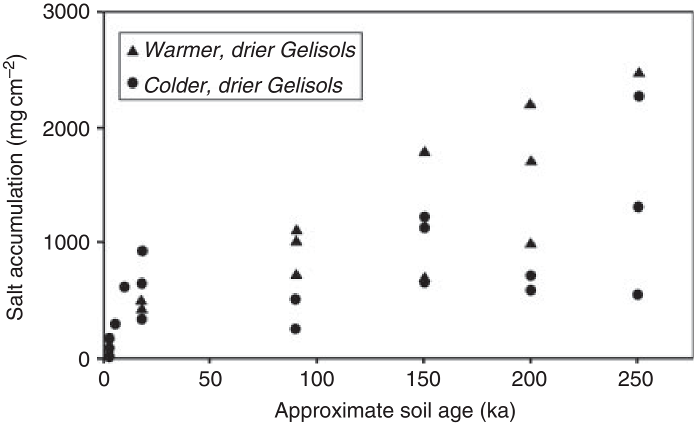

Figure 12.6. Relationship between total salt accumulation to a 70-cm depth and approximate soil age in Antarctica. Data are for relatively drier, colder (ultraxerous) and moister, warmer (xerous) Gelisols. Adapted from Bockheim (1997).

by winds in Gelisol landscapes. In Antarctica, the accumulation of soluble salts in soils is much faster than in the Arctic region. Gelisols in Antarctica receive significant amounts of salts primarily through atmospheric deposition; expression of salt accumulation ranges from encrustations beneath rock fragments to indurated pans (Bockheim 1997). Due to limited leaching, total salt content generally increases with time and can be used as an indicator of soil age (Figure 12.6). Pliocene-aged soils have salt-cemented pans that have formed from gradual, long-term input of NaCl or $NaNO_3$ from marine aerosols. Such soils are classified in salic and nitric subgroups (Bockheim and McLeod 2008; Soil Survey Staff 2010a).

Uses of Gelisols

Although Gelisols support the lowest population density of any of the soil orders, they are not immune to human impacts. Gelisols occur in areas harboring vast energy reserves, such as Alaska, and more than 500,000 people live in cities in Siberia that are built on permafrost (Bockheim and Tarnocai 2000). It is clear that Gelisol landscapes have very little resistance or resilience to disturbance, and they have proven to be fragile and extremely slow to recover.

One major impact of human disturbance on Gelisol landscapes occurs as heat flux into the soil is increased. This often results from disturbance or removal of insulating vegetative cover. In general, Gelisols having high contents of sand and gravel do not experience much change in strength or stability when permafrost thaws; however, finer-textured soils with disseminated ice can become supersaturated and lose all strength unless melt water can be drained (Moore 2004). This process can result in slope failure and considerable mass movement. *Solifluction* is a term for

Figure 12.7. Collapsed apartment building in Cherskii, northern Siberia. Sections of the building collapsed as underlying permafrost melted. (Image courtesy of Professor V. E. Romanovsky, University of Alaska, Fairbanks)

the slow flow of saturated soil and other unconsolidated materials. It is most rapid on steeper slopes and may move materials downslope at rates ranging from 0.6–1.5 m yr^{-1} (Tedrow 1977).

Because of the irregular occurrence of large ice blocks, wedges, or lenses, both above and below the permafrost table, melting of ground ice as a result of removal of a cover of peat and vegetation can result in hilly terrain where level land previously existed. The process of landscape collapse produces a new terrain called *thermokarst*. Melting of ice masses in Gelisols results in formation of thermokarst landscapes and disruption of most land uses. In the Siberian cities of Yakutsk and Norilsk, it is estimated that several hundred apartment buildings have been damaged as a result of melting permafrost (Goldman 2002). An example of this type of urban damage is shown in Figure 12.7. The process of cryoturbation must also be considered in Gelisol land use. Frost heaving, churning, and sorting can easily damage structures and roads if adequate engineering precautions are not taken.

Minimizing soil temperature changes can reduce human impacts on Gelisol landscapes. In building construction, structures can be placed on pilings or stilts to allow an air gap (~1 m) between the structure and the soil surface. This minimizes heat transfer to the soil and subsequent melting of permafrost. Placement of thick gravel pads on which buildings and roads are built has been used for petroleum operations infrastructure at Prudhoe Bay, Alaska. These 1.25-m-thick pads afford physical protection to the underlying soils and provide a thermal buffer, but reclamation of

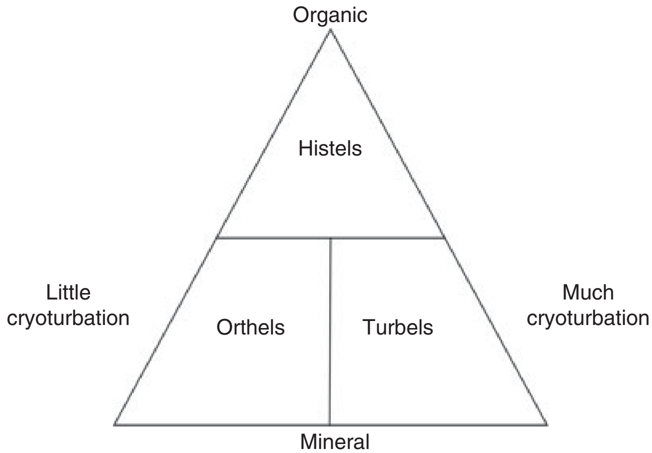

Figure 12.8. Diagram showing relationships among Gelisol suborders.

these gravel pads has proven to be challenging. An alternative to these engineering practices is to thaw permafrost prior to construction; however, this has the disadvantage of delaying projects for as much as five years (Goldman 2002).

Increased heat flux into Gelisols can also occur more indirectly. Construction and human traffic near McMurdo Station and Scott Base in Antarctica spread dark-colored dust over stable snow and ice, causing them to melt on sunny days. Severe water erosion is now actively altering the landscape during the warm season, and these buildings could eventually be lost.

Classification of Gelisols

Much of the basis for classification of Gelisols stems from work done by Russian, Canadian, and American soil scientists (Mazhitova et al. 1992; Agriculture Canada Expert Committee on Soil Survey 1987; Bockheim et al. 1993). In simplest terms, Gelisols are those soils that contain permafrost within 100 cm of the soil surface or have gelic materials within 100 cm of the soil surface and permafrost within 200 cm of the soil surface (Soil Survey Staff 2010a). Gelic materials are defined as "*mineral or soil materials that show evidence of cryoturbation (frost churning) and/or ice segregation in the active layer (seasonal thaw layer) and/or the upper part of the permafrost*" (Soil Survey Staff 2010a). Figure 12.8 illustrates the further differentiation of Gelisols into three suborders based on some soil properties and genetic factors.

Histels are Gelisols consisting of organic soil materials. Large quantities of organic carbon typically accumulate in these soils under anaerobic conditions or the organic matter present at least partially fills voids in pumiceous, cindery, or fragmental materials (Soil Survey Staff 1999). These soils occur predominantly in the Subarctic and Low Arctic vegetation regions of continuous or widespread permafrost (Bockheim et al. 1993). Histels are rare in Antarctica where cold desert conditions prevail. Histels

Table 12.2. Suborders and great groups in the Gelisol order

Suborder	Great Group
Histels	Folistels are saturated for <30 cumulative days unless artificially drained and are not artificially drained.
	Glacistels are saturated for ≥30 days cumulative days, have a glacic layer, and have <75% *Sphagnum* fibers.
	Fibristels are dominated by fibric materials.
	Hemistels are dominated by hemic materials.
	Sapristels—other Histels.
Turbels	Histoturbels have 40% or more organic soil materials.
	Aquiturbels have redox features within 50 cm of the mineral soil surface and aquic conditions (unless artificially drained).
	Anhyturbels have anhydrous conditions (dry permafrost).
	Molliturbels have a mollic epipedon.
	Umbriturbels have an umbric epipedon.
	Psammoturbels have a texture of loamy fine sand or coarser amd <35% rock fragments.
	Haploturbels—other Turbels.
Orthels	Historthels have 40% or more organic soil materials.
	Aquorthels have redox features within 50 cm of the mineral soil surface and aquic conditions (unless artificially drained).
	Anhyorthels have anhydrous conditions (dry permafrost).
	Mollorthels have a mollic epipedon.
	Umbrorthels have an umbric epipedon.
	Argiorthels have an argillic horizon within 100 cm of the soil surface.
	Psammorthels have a texture of loamy fine sand or coarser amd <35% rock fragments.
	Haplorthels—other Orthels.

are further subdivided into five great groups, where the formative element *hist* changes to *ist* (Table 12.2). Because of the slowness of decomposition of organic materials in Gelisol landscapes, Folistels and Fibristels are probably more common than the soils of the other two great groups.

Turbels (from the Latin, *turbides,* "disturbed") are other Gelisols that are characterized by evidence of cryoturbation. These soils contain one or more horizons with irregular, broken, or distorted horizon boundaries, ice or sand wedges, oriented stones, or other evidence of frost mixing (Soil Survey Staff 1999; Bockheim et al. 1993). Turbels occur primarily in the zone of continuous permafrost and are the second most-extensive suborder, accounting for ~43% of the Gelisols on a global basis (USDA-NRCS database). These soils are common in the High and Middle Arctic vegetation regions of North America and Eurasia at latitudes ≥65°N. Turbels also comprise approximately 50% of the McMurdo Dry Valleys of Antarctica (Bockheim and McLeod 2008). Turbels are subdivided into seven great groups (Table 12.2) on the basis of content of organic matter, natural drainage condition, soil climate, presence of diagnostic soil horizons, and presence of coarse textures.

Orthels (from the Greek, *orthos,* “true”) are the most extensive suborder and accounting for ~48% of Gelisols on a worldwide basis. Little or no cryoturbation occurs in these soils, and patterned ground (except for polygons) generally is lacking. These soils occur primarily within the zone of discontinuous permafrost, in alpine areas where precipitation is greater than 1,400 mm yr^{-1} (Burns 1990) and on coarse-textured materials at latitudes as great as 70°N in the High Arctic (Naumov et al. 1964; Kuzmin and Sazonov 1965; Ugolini 1986; Ugolini et al. 1981). Orthels are differentiated into eight great groups (Table 12.2) on the basis of many of the same criteria used in the differentiation of Turbels into great groups. In Antarctica, Anhyorthels are the dominant soil of the McMurdo Dry Valleys; these soils are characterized by dry permafrost and are typically loose, skeletal, and coarse textured (Bockheim and McLeod 2008). Data for one of these soils is provided in Table 12.1.

Perspective

Gelisols are soils that contain permafrost, thus distinguishing them from soils or the other orders that tend to be defined on the basis of chemical characteristics. Like the order of Vertisols, the Gelisols comprise landscapes that are seasonally unstable. Pedoturbation and some sort of patterning of ground are associated with both orders. Occurring primarily in polar and circumpolar regions, Gelisols occupy some of the harshest environments on earth and support only ~0.4% of the earth’s population. Because of cold temperatures, soil-forming processes operate very slowly and are often weakly expressed. In dry valleys of Antarctica with very low temperatures (0 to −70°C) and low precipitation (only traces of snow), Gelisols are unvegetated, quite inactive, and largely featureless even though they may be millions of years old. Warmer and moister Gelisols generally support tundra and boreal forest vegetation, contain significant quantities of organic matter, and show effects of cryoturbation. Careful management is required to sustain Gelisol landscapes, which are sensitive to disturbance. Wildfire, various human activities, and climate change can cause downward retreat of permafrost, creating thermokarst and potentially releasing carbon-based greenhouse gases.

Histosols: Organic Soils

13

Histosols (from the Greek, histos, tissue, and the Latin, solum, soil) is the name given in Soil Taxonomy to soils composed mainly of organic soil materials, do not have permafrost, and do not have andic properties dominant in the upper 60 cm of the soil. Organic soil materials are saturated with water for at least 30 days or are artificially drained and contain ≥12 to 18% organic carbon by weight, excluding live roots, depending on clay content. (See Figure 2.3.) If saturated for fewer than 30 days, organic soil materials must contain 20% or more organic carbon (Soil Survey Staff 1999). Histosols occupy approximately 1% of the land area (Wilding 2000). Many organic soils with permafrost that were previously classified as Histosols are now Histels, a suborder of the Gelisols occupying about 0.8% of the earth's land area.

Setting

Histosols occur at all latitudes, but about 90% of the Histosols occur in the boreal zone of North America, northern Europe (especially Finland and Sweden), Canada, and Russia (Everett 1983; Rabenhorst and Swanson 2000; Lindbo and Kozlowski 2006). Less extensive areas of Histosols are present in some lowlands throughout the tropics especially in Asia (Andriesse 1974). In this regard, the global distribution of Histosols is similar to that of Spodosols (Chapter 17). In the U.S. outside of Alaska, Histosols are locally extensive on the coastal plains of the Southeast, in the upper Great Lakes region, in southern Florida, and in central California near the confluence of the Sacramento and San Joaquin rivers. Elsewhere, Histosols occur only locally, where landscapes are very poorly drained or in alpine settings where cold temperatures retard organic matter decomposition. (See Figure 13.1.)

A general condition that must be met in the Histosols is that the rate of organic matter production must exceed the rate of organic matter decomposition. This condition may be met when water tables are maintained very near the soil surface for most of the year. As a result, less oxygen is available for aerobic microbial decomposition of organic residues. The anaerobic respiration process is less efficient than aerobic decomposition, organic substrates are less completely metabolized, and microbially synthesized biomolecules may be resistant to further oxidation (Anderson 1995). Thus, Histosols are most common in climates where precipitation exceeds evapotranspiration. Cool climates promote this process because of reduced microbial

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Figure 13.1. An alpine Cryohemist in Switzerland, formed mostly from sedges, with gray strata of mineral soil material. The black bar represents 50 cm. The water table is at a depth of about 80 cm. For color detail, please see color plate section.

decomposition, hence, the increased presence of organic matter in freezing conditions and low evaporation. The accumulation of peats in a wetland area generally depends more strongly on the decomposition rate than on the biomass production rate; decomposition rates are more a function of temperature than of precipitation (Mausbach and Richardson 1994).

A variety of terms are used to describe organic soils or landscapes dominated by them (Stanek and Worley 1983) including *mire, moor, bog, peatland, muskeg* (Canadian Algonquin term), *pocosin* (Carolinian term), *fen, marsh,* and *swamp*. These names apply to particular ecosystems or landscapes characterized by specific biotic communities, hydrologic regimes, reaction (pH), nutrient status, and pattern of microrelief and ponds. For example, bogs are nutrient poor, have acidic organic soils, and are dominantly rain fed. Fens or swamps (if forested) have less acidic organic soils, are more nutrient rich, and are fed dominantly by groundwater or runoff. Histosols also occur in subaqueous settings, for example, in intertidal areas, where the soils are submerged most of the time (Soil Survey Staff 2010).

Hydrology clearly is a very important factor in the formation and maintenance of Histosols and can be considered as a component of both climate and topography (Mausbach and Richardson 1994). Histosols may occupy a variety of landscape positions (Figure 13.2), and their properties depend largely on the hydrologic regime. Histosols often occur on the lowest, wettest parts of the landscape. Those Histosols that occupy landscape depressions are typically independent of climate ("aclimatic") and are the result of a high water table. Some Histosols are formed on hillslope seep

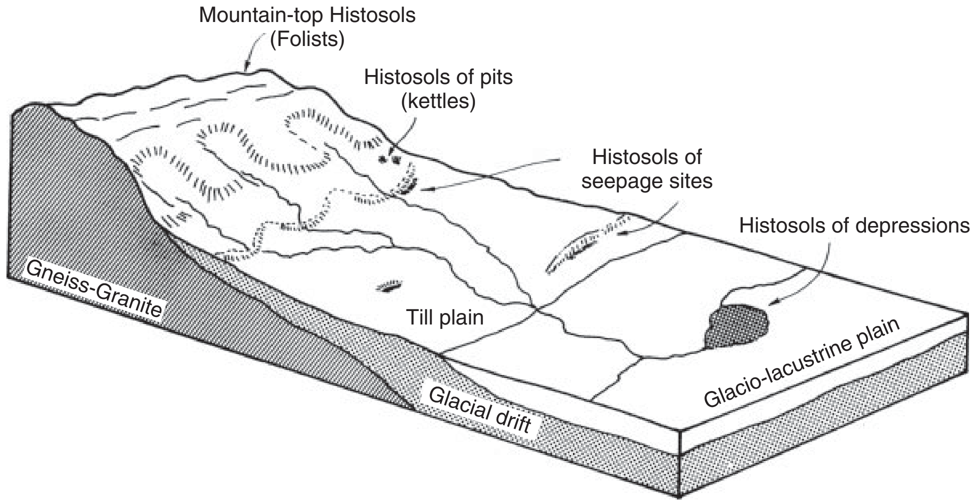

Figure 13.2. Idealized block diagram showing some relationships of Histosols to topographic position.

areas due to local stratigraphy that causes lateral groundwater movement and local seepage on side slopes. Blanket peats and raised bogs depend on rainfall and often occupy the highest part of a poorly drained, low relief landscape. Fens often occur where groundwater discharge or stream inflow supplies water and nutrients. In perhumid climates, where precipitation exceeds evapotranspiration several months each year, Histosols form on flat uplands that occur on the centers of interstream divides and are known as "pocosins" (pocosin—American Indian word for "swamp on a hill"). Pocosins (bogs) are common in the lower coastal plain of North Carolina and actually are "peat domes" formed in drainage systems that were blocked and flooded between 10,000 and 15,000 years ago. Some Histosols of cool, humid mountainous regions ("Mountain-top Histosols" in Figure 13.2) are never saturated with water, except for a few days following heavy rain. They are either shallow or extremely rocky, and the plant roots grow only in the organic material.

Bodies of eutrophic Histosols (fens) are chemically influenced by input of nutrients (Verry 1981). Houghton muck (Figure 13.3) receives nutrients from stream and seepage water moving from adjacent slightly calcareous sandy glacial till. By contrast, the rain-fed, infertile raised bog, classified as Napoleon, receives nutrients only from precipitation and wind.

The parent materials for most Histosols are hydrophytic plants. Mosses of the genus *Sphagnum* dominate many Histosol landscapes. Other plant species commonly providing organic material to Histosols include sedges (*Carex* spp.), rushes (*Juncus* spp.), and cattails or tules (*Typha* spp. *and Scirpus* spp.). Ericaceous shrubs are common in some slightly better drained Histosol landscapes, as are stunted trees including spruce (*Picea*), pine (*Pinus*), hemlock (*Tsuga*), and willow (*Salix*). Just as rock composition affects soil properties of mineral soil, plant composition affects the

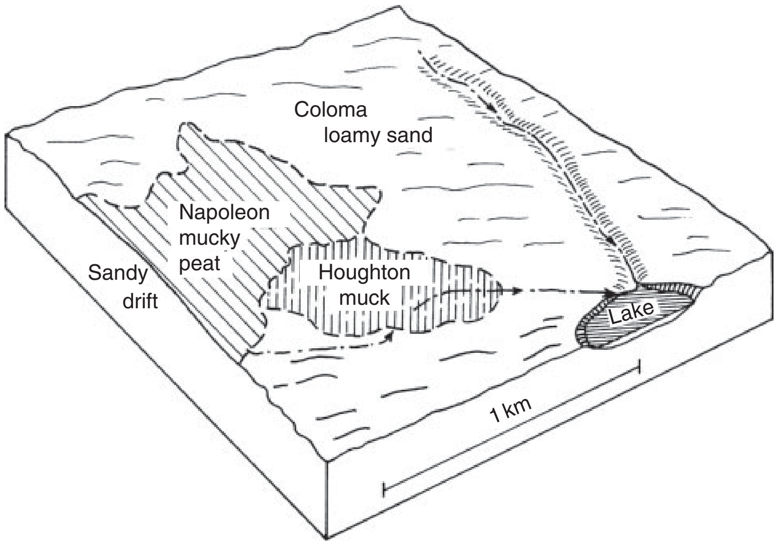

Figure 13.3. Soilscape patterns of two Histosols, in association with Entisols in slightly calcareous till terrain in Van Buren County, Michigan. Coloma soils are mixed, mesic Lamellic Udipsamments. Napoleon soils are dysic, mesic Typic Haplohemists. Houghton soils are euic, mesic Typic Haplosaprists. (After Bowman 1986, Sec. 3, T.1S., R.15W)

properties of organic soils. Three broad classes of peat are identified on the basis of plant composition (Rabenhorst and Swanson 2000): moss peat (generally in bogs), sedge or herbaceous peat (generally in fens), and woody peat (generally in swamps).

The Histosols of the boreal zone occupy landscapes that were mostly covered by ice during the Pleistocene. These Histosol landscapes must be less than about 10,000 years old. At lower latitudes, most Histosols of coastal plains and estuaries probably have ages of 5,000 years or less. These Histosols probably began to form after sea level more or less stabilized at its current high stand following melting of ice during the late Pleistocene and early Holocene (Rabenhorst and Swanson 2000). Locally, Histosols may be considerably older than or younger than these general age ranges due to local hydrologic and topographic conditions.

Geological Processes of Organic Matter Accumulation

The geologic accumulation of organic materials and the extension of "blanket peats" over entire landscapes are termed paludization or paludification (Malmer 1975; Glaser 1987; Mausbach and Richardson 1994). The enlargement of such bodies is by surface additions of organic materials. Anoxic conditions created by prolonged saturation, reducing the infusion of oxygen needed to oxidize the organic material, must predominate at the soil surface for paludization to occur. The slow hydraulic conductivity of many Histosols may play a positive feedback role in the paludization process. Slow hydraulic conductivity leads to longer periods of saturation and anaerobic conditions, resulting in retarded organic matter decomposition, hence, more organic matter accumulates, leading to slow hydraulic conductivity (Hartshorn et al. 2003). Paludization can be a final stage in another process called "terrestrialization" (Malmer 1975), the in-filling of lakes and other depressions with sediment, and production of the classic "peat land" (Glaser 1987). These wetlands are well suited for Histosols formation because of the low organic matter decomposition rates under anaerobic conditions in the saturated soil. Paludization forms the blanket bogs on the mineral soil surface of a forested or tundra terrain and proceeds from a muskeg stage to a raised bog stage (Henselman 1970).

In Vermont, the biomass of *Sphagnum* moss has been observed (Osheyack and Worley 1981) to increase 16% annually for an average areal production of $370 \pm 330\,g\ m^{-2}$ dry matter. This annual production rate is one-half the rate of boreal forests, one-third that of north temperate forests, one-quarter that of tropical forests, but 1.3 times that of alpine heath, 2.2 times that of alpine tundra meadow, 4 times that of deserts, 1.5 times that of streams, and 3 times that of the open ocean. Mined bodies of Histosols in Scotland are reported (Robertson 1981) to regrow ("heal") within 5–10 years, but no such healing has been noted in Minnesota where *Typha* spp. (cattail, with biomass production of 33 tonnes $ha^{-1}\ yr^{-1}$) has been tried as a bioenergy crop following removal of peat (Garver et al. 1983). In some Histosols of the Florida Everglades (McDowell et al. 1969), the organic material a few centimeters above a limestone contact is about 4,300 years old, and the material 1.26 meters above the limestone is about 1,250 years old.

Since genesis of Histosols depends on organic matter deposition, the deposition process is often considered to be geogenic rather than pedogenic. In this sense, one can consider the initial deposit of organic materials to be the "parent material" in which Histosols can form by alteration from recognizable organic forms of leaves, stems, and other plant parts, to unrecognizable organic material; and from a stratified or unstructured mass to granular, blocky, or prismatic structured horizons.

Pedogenic Biogeochemical Processes

The decomposition of organic matter is controlled by a number of interrelated factors of which moisture content, temperature, composition of the deposit, acidity, microbial activity, and time are the most important (Broadbent 1962). The alterations and reactions taking place during decomposition are complicated and only partially understood. The decomposition, transformation, and physical alteration of the initial organic material to produce Histosols are often referred to as ripening. In the Netherlands, Heuvelen et al. (1960), Jongerius and Pons (1962), and Pons (1960) have considered ripening processes of Histosols to begin as soon as microbial activity is promoted by entry of air into the organic deposit. Physical ripening primarily involves a decrease in volume due to consolidation and loss of material via decomposition. The amount of physical ripening depends upon the nature of plant remains, the content of mineral matter, and the depth of the water table. Chemical ripening is a combination of the chemical decomposition of the most labile components, the partial metabolism of more resistant components leaving still more resistant remnants, and the biosynthesis of new compounds as part of the microbial biomass (Kononova 1961; Anderson 1995). The chemical processes also include transformations of organic and inorganic sulfides, especially in coastal environments where brackish water provides dissolved sulfates (Rabenhorst and Swanson 2000). If the saturating conditions are terminated in sulfide-rich soils, often by engineered drainage systems, oxidation of sulfides produces sulfuric acid, which strongly acidifies the soils, and may lead to the formation of a sulfuric horizon (pH 3.5 or less and jarosite, underlying sulfidic materials, or at least 0.05% water-soluble sulfates). Biological ripening involves reduction in particle size, mixing of organic materials, and formation of peds and pedological features by organisms.

The first organic substrates to be metabolized by microorganisms, primarily fungi and bacteria, are the relatively simple biomolecules including amino acids, soluble proteins, and simple carbohydrates (Anderson 1995; Everett 1983). The products of this decomposition are CO_2 and water, plus energy, C, and N for the synthesis of microbial biomass. More complex organic molecules including hemicellulose, cellulose, and lignin are metabolized less easily and yield a complex mixture of long aliphatic (polymethylene) chains, and aromatic compounds (containing phenol groups). This mixture is a combination of recalcitrant remnants of the initial organic substrates, plus microbially synthesized material (Anderson 1995). Anaerobic conditions prevent the complete oxidation of the reduced carbon in the organic molecules to CO_2.

The decomposition process has a profound effect on the physical and chemical properties of organic soil materials (Table 13.1). As the organic molecules are decomposed and transformed, C is lost (e.g., cellulose), but N is conserved through microbial synthesis. The C:N ratio decreases as decomposition proceeds. The cation exchange capacity (CEC) of Histosols is derived primarily from carboxyl and phenolic functional groups (Broadbent and Bradford 1952). The number of these functional groups increases as decomposition progresses, and CEC values of 200 cmol kg^{-1} and higher have been reported for some organic matter (Broadbent 1953). Note that the effect of decomposition on CEC is even greater when CEC is reported on a soil volume basis, rather than on a mass basis, due to the change in bulk density. The CEC of many organic soils is pH-dependent and may change from about 10 cmol kg^{-1} at pH 3.7 (in 1:1 water) to over 100 cmol kg^{-1} at pH 7 (Dolman and Buol 1967).

Most Histosols have bulk densities much less than 1 g cm^{-3}. Bulk density tends to increase with decomposition (Boelter 1965) from values as low as 0.02 g/cm^3 in fibric materials to 0.3 g/cm^3 or more in sapric materials as plant structural parts are broken down and voids collapse. The amount of mineral material and the type of vegetation contributing to the Histosol causes a wide variation in the bulk density. Histosols are commonly saturated and have an extremely high water-holding capacity, both on a weight

Table 13.1. Selected representative or mean values for properties of organic soil materials in relation to degree of decomposition

Property	Slightly Decomposed (Fibric)	Moderately Decomposed (Hemic)	Highly Decomposed (Sapric)
Organic carbon (%)	37	35	31
Total nitrogen (%)	1.4	1.6	1.8
C:N	28	25	21
Cellulose (% dry weight)	23	22	16
pH	4.5	4.9	5.1
CEC_7 ($cmol_c$/kg)	83	88	101
CEC_7 ($cmol_c$/L)	21	44	76
Bulk density, field moist (g/cm^3)	0.02–0.08	0.07–0.18	0.10–0.30
Water content (% by volume, 10 kPa)	20	50	75
Water content (% by volume, 1500 kPa)	10	15	25
Vertical hydraulic conductivity (cm/s)	0.045	0.00056	0.00027
Horizontal hydraulic conductivity (cm/s)	0.089	No data	0.00027

Source: Rabenhorst and Swanson 2000; Everett 1983.

and a volume basis. Much of the water is either in the larger pores (gravitational water) or in such small pores that it is unavailable for plant growth (Boelter and Blake 1964). In general, although the least decomposed organic soil materials have the greatest water content at saturation, they tend to have the lowest plant available water due to the large proportion of very large pores that empty at low suction. Because Histosols shrink considerably on drying, their moisture characteristics are best expressed on a wet bulk volume basis or measured in the field (Boelter 1964).

Hydraulic conductivity decreases significantly as decomposition proceeds from fibric through hemic to sapric materials. In many Histosols, the degree of decomposition increases with increasing depth, thus hydraulic conductivity decreases with depth, in many cases by two or three orders of magnitude (Hartshorn et al. 2003). Also, horizontal hydraulic conductivity may be two or three times greater than vertical hydraulic conductivity. These conditions favor slow vertical movement of water and significant lateral flow of water through and out of Histosol landscapes.

Uses of Histosols

Histosols are used for production of crops, trees, wildlife, and recreation. Organic material is harvested for (1) horticultural potting materials, (2) chemicals (extractable), (3) fuel for heating and for generation of electricity, and (4) treatment of wastewater and sewage.

Management of Histosols for growing crops (including trees; Berguson et al. 1983) involves drainage (Stephens 1955; Roe 1936), irrigation when appropriate, fertilization, cultivation (Farnham 1983), and several special harvesting practices.

The Houghton soils of Figure 13.2 are suitable for production of carrots, onions, potatoes, mint, and salad crops. The low-bulk density of Histosols permits root crops to attain perfect shapes. Forest production improves with drainage of Histosols (Maki 1974). Drainage firms the soil enough to support field equipment. However, dewatering leads to oxidation of the soil by volatilization, chiefly as water and carbon dioxide, and to further subsidence due to consolidation. The oxidation process does release plant nutrients from the organic materials, but far too slowly to satisfy needs of growing plants. Hence, careful fertilization of crops on Histosols must include not only macronutrients, but also micronutrients, such as copper, according to soil tests. Bodies of Histosols are typically in depressions, which, in temperate zones, may have cold microclimates because of cold air drainage to the lowlands. In cranberry bogs, when night frosts are predicted, spray irrigation and even flooding are used to save the crop in these landscape positions with short growing seasons.

Subsidence in Florida was found by Stephens (1956) to be about 3 cm per year under cultivation. The rate of subsidence was directly related to depth of artificial drainage and could be predicted by the following formula: $x = (y - 2.45)/14.77$, where x is the annual subsidence in inches and y is the average depth of drainage in inches. The subsidence rate in Florida was about twice as fast as subsidence of drained Histosols in Indiana where the formula $x = (y - 9.6)/23$ was applicable. Undoubtedly, the difference is related to higher temperatures and lack of freezing in Florida. Another study in Florida (McDowell et al. 1969) indicated that 1.8 meters of a Histosol were lost after 50 years of drainage, or according to the researchers' calculations, about as much as had been created in 1,200 years. Therefore, a description of Histosols is incomplete without reference to their susceptibility to subsidence. As a result of subsidence, soil maps of Histosols in Florida's Everglades (Collins et al. 1986) must be updated regularly. Ground-penetrating radar (Baraniak 1983) is useful in making continuous recordings of thickness of Histosols along transects. In the agricultural part of the Everglades, 66% of the acreage of Lauderhill muck (euic, hyperthermic Lithic Haplosaprists) had thinned during the period 1974–1984 to Dania muck (euic, hyperthermic, shallow Lithic Haplosaprists). It is estimated that between the years 1914 and 2000, soils of 90,000 ha of the original 250,000 ha of Histosols used for agricultural production in this region decreased in average thickness from 3.6 m to 30 cm.

The hydrologic characteristics of Histosols are closely related to the degree of decomposition (Table 13.1). In particular, highly decomposed materials (sapric) may have hydraulic conductivities lower than those for clay-textured mineral sediment (Boelter 1965), so low that open ditches cannot effectively control the water table unless very close spacing is used (Boelter 1974). An effective system for drainage of Haplosaprists with extremely low hydraulic conductivity is to space the ditches about 100 m apart and shape the fields between ditches to a relief of about 25 cm, providing for surface runoff and lateral flow to the ditches.

Fire hazard increases after drainage of Histosols. Peat and muck fires are difficult to control and may burn for several months, polluting the air, as well as destroying the organic material. Once the organic soil has been destroyed by fire, farming may continue on the remaining mineral substratum, which is usually poorly drained (Davis and Engberg 1955; Jongedyk et al. 1950, 1954; Mirza and Irwin 1964; Neller 1944). In some areas, resistant, hard logs and stumps concentrate on the surface as the other organic materials oxidize. These create difficulties for the operation of equipment (Dolman and Buol 1967). Decomposition and drying of surface organic material make the soil very susceptible to wind erosion due to low bulk density and low cohesion.

Building structures and roads on Histosols is usually difficult. The organic material has very low bearing capacity, and most structures must be placed on foundations that extend to underlying mineral material. If an area is drained just prior to time of construction, subsidence of the soils becomes apparent only gradually. Ultimately the structures are left standing far above the ground level. In the case of house construction on piles driven into the mineral soil, the building becomes a "house on posts" perched above the lawn. The garage stands above the driveway and is inaccessible to the car (Slusher et al. 1974). Peatland (Histosol) waters are known to corrode both concrete and metal (MacFarlane and Williams 1974).

Horticultural value of Histosol materials lies in their use in potting soil, as conditioners for mineral soil, as plant stimulants, and as a component of some fertilizers. A vast array of chemicals (including dopplerite, a calcium salt of a humic acid of commercial and medicinal value) has formed in Histosols (Gary et al. 1983; Pihlaja et al. 1983). Useful waxes have been separated from peat in Finland (Peltola et al. 1983).

Histosols and their parent materials are used for fuel, either in burning directly or after conversion to methane by gasification (LeMasters et al. 1983). Technology is well advanced for mining, dewatering, and burning organic materials to heat buildings and to generate electricity (Farnham 1978; Chornet et al. 1982; Dubbe et al. 1982). Peat is used extensively as a fuel in northern Europe especially Belarus, Russia, Finland, and Ireland (Rabenhorst and Swanson 2000). Comparative energy values of fuels, in BTU lb^{-1}, dry, are as follows: wood, 6,000; grasses and sedges, 7,500; fibric soil, 7,300; sapric soil, 8,300; lignite, 9,500; anthracite, 14,000.

"Peatlands" are being recognized more and more as important sanctuaries for unique plant communities (insectivorous species, orchids, tundra species) and associated animals. For many years, peatlands were avoided and ignored, or were viewed by many people as wastelands, that were of value only if drained and cultivated. We now recognize the value of these lands as components of wetlands that provide habitat for plants and animals, act as filters of N, P, sediment, and other contaminants (Farnham 1974; Nichols 1981; Loxham and Burghardt 1983), and, while forming, act as sinks of atmospheric carbon (Hartshorn et al. 2003). Undisturbed bodies of Histosols hold scientific interest and aesthetic value (including "the heath's fragrance," mentioned by McDonald [1981]) that attract visitors in their leisure time. "Scientific areas" have been set aside in several states of the United States as gene pools, research areas, and nature appreciation landscapes.

Describing Histosols

Conventions useful in describing mineral soils, as shown in Chapter 2, are not wholly applicable to Histosols. Each layer of a Histosol is described in the following terms: *color, fiber content before and after rubbing, structure, consistence, roots, reaction, boundary,* and *color of the sodium pyrophosphate extract* (Soil Survey Staff 1999). Moist and dry colors are determined, when possible, and in addition, the "rubbed" color of the material is determined after it has been pressed or rubbed in the hands. Unrubbed fiber content is an estimate of the volume of plant fibers observable in a freshly broken surface. Rubbed fiber content is estimated after the material has been rubbed between the fingers to break down extremely rotten fibers. The estimate of fiber content excludes live roots. Structure, consistence, content of live roots, reaction, and boundaries are described as in mineral soils. Sodium pyrophosphate extract color (Munsell) is measured after absorbing solution from a mixture of soil material and sodium pyrophosphate on white filter paper (Kaila 1956; Soil Survey Staff 1996).

Additional information of importance may include the botanical origin of the fibers, notes on included thin strata of organic or mineral soil material, presence of logs and stumps, mineral content, and presence of sulfate minerals (e.g., jarosite) or sulfidic material (sulfide-containing material that becomes very acidic upon oxidation due to sulfuric acid production). All depths are recorded from the soil–air interface.

Kinds of Organic Soil Materials. Three kinds of organic soil materials—fibric, hemic, and sapric—are recognized on the basis of the extent of decomposition of the plant material from which they are formed (Soil Survey Staff 1999).

Before defining and describing these three soil materials, it is necessary first to describe fibers as they occur in organic soils. Fibers are short pieces of plant tissue in organic soil materials (except for live roots) that (1) are large enough to be retained on a 100-mesh sieve when the materials are sieved after sodium hexametaphosphate dispersion; (2) show cellular structure of the source plants from which they are derived; and (3) must be 2 cm or less in their smallest dimension, or are decomposed enough so they can be crushed with the fingers.

Fibric soil materials are defined as organic soil materials that either (1) contain three-fourths or more by volume of fibers after rubbing, or (2) contain two-fifths or more by volume of fibers after rubbing and yield a sodium pyrophosphate extract with color value and chroma of 7/1, 7/2, 8/1, 8/2, or 8/3. Fibric materials are generally lighter in color, and browner or redder than more decomposed organic soil materials.

Hemic soil materials are between the less decomposed fibric and the more decomposed sapric materials.

Sapric soil materials are the most highly decomposed of the three kinds of organic soil materials. They have a rubbed fiber content of less than one-sixth by volume and yield sodium pyrophosphate extracts with Munsell color value less than 8, and with value and chroma that are below or to the right of a line drawn to exclude value/chroma of 5/1, 6/2, and 7/3. They are mostly dark gray to black, have the smallest amount of plant fiber, highest bulk density, and lowest water content (on dry weight basis) at saturation.

In addition to fibric, hemic, and sapric materials, humilluvic material is also recognized.

Humilluvic material (illuvial humus) accumulates in lower layers of some Histosols (usually old, drained, and cultivated). The humilluvic material is highly soluble in sodium pyrophosphate and commonly accumulates at a contact with sandy mineral soil material. To be recognized in classifying Histosols, humilluvic material must make up one-half or more (volume) of a soil layer 2-cm or more thick. Hydraulic conductivity of humilluvic material is very slow.

Limnic materials, present in some Histosols, include both organic and inorganic materials that were either deposited in water by algae or diatoms or originated from underwater and floating aquatic plants and were modified by aquatic animals. *Coprogenous earth* (sedimentary peat layer) is a limnic layer that has many fecal pellets, Munsell color value of 4 or less, forms viscous water suspensions, and is nonplastic or slightly plastic or shrinks upon drying and is difficult to rewet, and either has a saturated sodium pyrophosphate extract Munsell value of 7 or more and chroma of 2 or less or has a CEC less than 240 cmol kg^{-1} of organic matter. *Diatomaceous earth* is a limnic layer composed mostly of diatoms that has a Munsell color value of 3, 4, or 5, if not previously dried, and yields a sodium pyrophosphate extract with a color value of 8 or more and chroma of 2 or less or has a CEC less than 240 cmol kg^{-1} of organic matter, or both. *Marl* is a limnic layer composed mostly of carbonates that has a moist color value of 5 or more, and effervesces with dilute hydrochloric acid.

Horizon nomenclature has been developed that relates directly to diagnostic layers (Chapter 2). Organic horizons are denoted by O; subhorizons are designated as i (Oi) for fibric, e (Oe) for hemic, and a (Oa) for sapric. Limnic layers are designated as L layers.

Von Post (1924) developed a 10-stage scale of degree of decomposition of organic soil materials, based on the proportion of material remaining in the hand after squeezing a wet sample. If colorless liquid is produced upon squeezing, it is little decomposed; if all of the organic material escapes between the fingers, the soil is classified at the highest stage of decomposition. With practice, one can get good results with this technique in identifying fibric, hemic, and sapric materials. Fibric material produces only slightly turbid water and no material protrudes between the fingers (Von Post stages 1–3). Hemic material produces turbid water, and less than two-thirds of the original handful escapes between the fingers upon squeezing; that is, more than one-third of the handful is retained (Von Post stages 4–7). Upon squeezing wet sapric material, more than two-thirds of the sample extrudes between the fingers (Von Post stages 8–10).

Classification of Histosols

Three important elements are of major concern in defining the order: (1) a standard minimum content of soil organic matter; (2) a required thickness of this organic soil material; and (3) avoidance of criteria that would necessitate reclassification of the soil as a result of common agricultural practices such as drainage.

Organic soil material, if saturated with water for more than 30 days or if artificially drained, has at least 18% organic carbon if the mineral fraction has 60% or more clay; at least 12% organic carbon if no clay is present; or proportionately more organic carbon than a line connecting these points for intermediate clay contents. (See Figure 2.3.) If the soil is saturated for fewer than 30 days, it must contain 20% or more organic carbon.

Generally, Histosols do not have permafrost and do not have andic properties in 60% or more of the upper 60 cm of the soil, and half or more of the upper 80 cm of soil is organic soil material. Histosols are also identified if the organic materials rest on rock or fill or partially fill voids in fragmental, cindery, or pumiceous materials. If the bulk density of the soil is less than 0.1 g cm^{-3}, then three-fourths or more of the upper 80 cm must be organic soil material.

An arbitrary control section of 130 cm or 160 cm is defined for use in classifying Histosols, providing no densic, lithic, or paralithic layer or water or permafrost occurs in that depth. The thicker limit is used only when the surface 60 cm is more than 75% moss fibers or has a bulk density less than 0.1 g cm^{-3}. This control section is then subdivided into three layers referred to as surface, subsurface, and bottom tiers. The surface tier is 30-cm thick unless 75% or more of it consists of moss fiber or has a bulk density less than 0.1 g cm^{-3}, in which case it is 60-cm thick. The subsurface tier is 60-cm thick and may include mineral material, provided it is not densic, lithic, or paralithic. The bottom tier is 40-cm thick or it extends to the top of a densic, lithic, or paralithic layer, or a layer of water or permafrost, whichever is shallower.

The Histosols order is divided into five suborders (Figure 13.4). Histosols that are saturated with water for fewer than 30 days per year are in the suborder Folists. Histosols that have positive water pressure at the surface for more than 21 hours each day are in the suborder Wassists. The Wassists were added as a Histosol suborder in the 11th edition of *Keys to Soil Taxonomy* (Soil Survey Staff 2010), to identify subaqueous soils dominated by organic soil materials. These subaqueous soils are common in the intertidal zone and adjacent areas submerged by water up to 2.5-m deep (Demas and Rabenhorst 1999; Bradley and Stolt 2003; Osher and Flannagan 2007). The other three suborders are differentiated on the basis of the degree of decomposition of the organic material, primarily in the subsurface tier. Fibrists have fibric material dominant in the subsurface tier and do not have a sulfuric horizon within 50 cm of the surface or sulfidic materials within 100 cm of the surface; Saprists have a subsurface tier dominated by sapric material; Hemists are other Histosols (that have a subsurface tier dominated by hemic organic material).

A listing of the great groups recognized in the Histosol order is given in Table 13.2. A variety of criteria are used to separate great groups. A cryic soil temperature regime is used in all of the suborders other than the Wassists. Soil moisture regime is used to separate the Folists because they are not saturated for long periods. Sulfidic materials and salinity are properties used to differentiate great groups

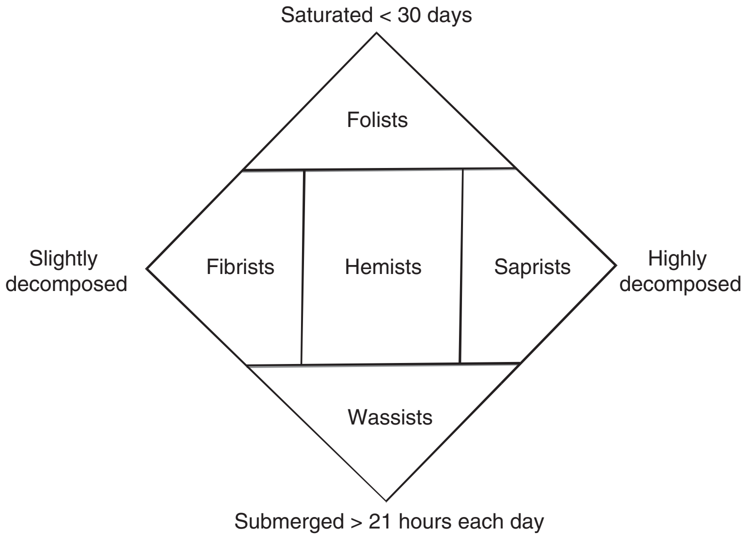

Figure 13.4. Diagram showing some relationships among suborders of Histosols.

Table 13.2. Suborders and Great Groups in the Histosols Order

Suborder	Great Group
Folists	Cryofolists have a cryic soil temperature regime.
	Torrifolists have an aridic soil moisture regime.
	Ustifolists have an ustic or xeric soil moisture regime.
	Udifolists—other Folists.
Wassists	Frasiwassists have electrical conductivity of less than 0.2 dS/m in all horizons within 100 cm.
	Sulfiwassists have sulfidic material within 50 cm.
	Haplowassists—other Wassists.
Fibrists	Cryofibrists have a cryic soil temperature regime
	Sphagnofibrists have *Sphagnum* as 75% or more of the volume to a depth of 90 cm, or to fragmental materials, a densic, lithic, or paralithic contact, or to a mineral soil contact if shallower.
	Haplofibrists—other Fibrists.
Saprists	Sulfosaprists have a sulfuric horizon within 50 cm.
	Sulfisaprists have sulfidic materials within 100 cm.
	Cryosaprists have a cryic soil temperature regime.
	Haplosaprists—other Saprists.
Hemists	Sulfohemists have a sulfuric horizon within 50 cm.
	Sulfihemists have sulfidic materials within 100 cm.
	Luvihemists have a 2-cm or thicker horizon containing 50% or more by volume humilluvic materials.
	Cryohemists have a cryic soil temperature regime.
	Haplohemists—other Hemists.

of the Wassists. Presence of moss fibers, humilluvic materials, sulfidic materials, and a sulfuric horizon are properties used to identify great groups in the other suborders. The nature of the bottom tier, including the presence of a mineral soil contact, is considered in the subgroup classification.

Perspective

Histosols form in surficial bodies of organic material produced mostly by plants and generally accumulate under saturated, anaerobic conditions that slow decomposition (Folists are the exception) or in subaqueous settings where organic soil materials are dominant. Generally, to qualify as a Histosol, the soil must not have permafrost or andic properties, must contain at least 12 to 18% organic carbon by weight if saturated with water, or at least 20% organic carbon if not saturated, and be at least 40-cm thick. Other properties important for classification include degree of decomposition, positive water pressure due to submersion, pH, presence of sulfides or their acidic products of oxidation, temperature, salinity, and depth to mineral soil materials. Decomposition causes loss of carbon, an increase in bulk density, water-holding capacity and cation exchange capacity, a lower C:N ratio, and slower hydraulic conductivity. Histosols are mined for fuel and for horticultural purposes. Intensive production of root vegetables, salad crops, sugarcane, and other special crops involves artificial drainage, irrigation, and fertilization. Local conditions of soilscapes and of economic and cultural conditions determine the uses for these soils. Some areas are left undisturbed to serve as wildlife refuges and as natural bodies for storage and filtering of water and for flood control. If drained, these soils oxidize steadily, thereby contributing to atmospheric CO_2 levels. They subside due to this aeration and to consolidation by dewatering, and are susceptible to fires and wind erosion. Long-term land-use plans are needed for managed peatlands to maintain the soil-forming conditions that result in a net accumulation of organic soil materials.

Inceptisols: Embryonic Soils with Few Diagnostic Features 14

Inceptisols are soils that have not developed features diagnostic for other orders but have some features in addition to the ochric epipedon and albic horizons permitted in the Entisols. Originally, such a wide diversity of soils was placed in the Inceptisol order that the prinicipal author of *Soil Taxonomy* (1975), G.D. Smith, referred to it as the “wastebasket order” (Smith 1986). As knowledge of soils around the world increased, the Andisols order (Chapter 9) was created primarily of soils that had been classified in the former Inceptisol suborder, Andepts. Many soils formerly classified as Cryaquepts, Cryochrepts, and Cryumbrepts with permafrost or gelic materials are now included in the Gelisol order (Chapter 12).

Setting

Inceptisols have profile features more weakly expressed than those of many other soils and retain a close resemblance to their parent material (Figure 14.1). By definition, Inceptisols cannot have an aridic soil moisture regime; thus, they are excluded from most arid regions. Many Inceptisols are on steeply sloping, often mountainous landscape positions throughout all but the arid regions of the world. In the colder areas of the temperate zone, Inceptisols are extensive in areas of glacial deposits that are only a few thousand years old. Nearly level, relatively recent valley and delta deposits of rivers such as the Mississippi, Ganges, Amazon, and many others are dominated by Inceptisols.

Because Inceptisols are developed in a variety of climates, excluding arid regions, the temperature and rainfall distribution cannot be discussed with specific reference to the order. As an example of this climatic diversity, representatives of the Xerept suborder in California have soil temperature regimes ranging from thermic to cryic and occur under mean annual precipitations of 250 to 2,750 mm. Dystrudepts in Taiwan have formed under mean annual rainfall of 4,800 mm (Chen et al. 2001).

Pedogenic Processes

No single pedogenic process operates in all Inceptisols except leaching. It is probably more correct to say that virtually all of the pedogenic processes are active to some extent in Inceptisol profiles but none predominates. In steep terrain, mass wasting

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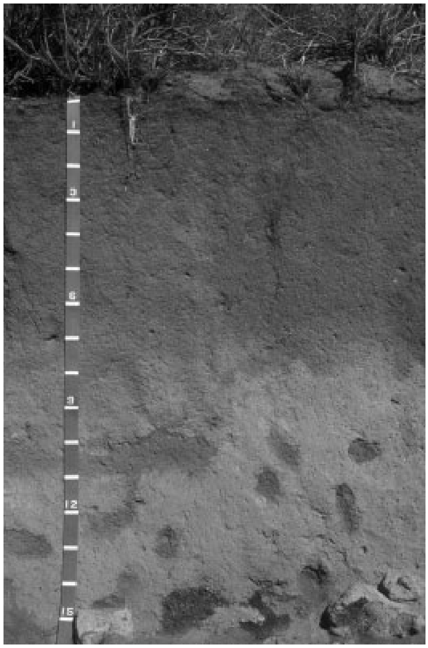


Figure 14.1. Photograph of a Vitrandic Haploxerept near Craters of the Moon National Monument and Preserve, Idaho. Parent materials are mixed ash/cinders and loess. The mixed parent material, young age (~2,000 years), and relatively cool, dry climate have impeded development of andic soil properties so the soil is an Inceptisol rather than an Andisol. Note the 5-cm-diameter krotovinas in the lower part of the profile filled with yellowish brown material from the upper horizons. This is evidence of mixing by burrowing rodents. Scale is marked in 10-cm increments. For color detail, please see color plate section.

processes remove soil material before the effects of pedogenic processes become strongly expressed (Jien et al. 2009). Most Inceptisols on steep slopes are in equilibrium with their present environment, and their properties will change little until that environment changes. Inceptisols on steep slopes with relatively rapid geologic erosion rates will no longer be there when geologic erosion has leveled the slope to a more subdued relief. On more nearly level surfaces, erosion rates are slower, and more distinct pedogenic features like argillic horizons are formed. Usually the pedogenic processes will create new, but similar, Inceptisol pedons in underlying volumes of parent material as erosion lowers the landscape by removing the volume of material that was soil. Deep colluvium at the base of steep slopes frequently results from landslides, but slow mass movement called soil creep (Selby 1982) and surface erosion also add material to such colluvium. In the Blue Ridge Front of North Carolina, the upper 10 to 20 cm of Inceptisols and Ultisols formed over mica schist and gneiss saprolite were found to be composed of material moved by creep and/or erosion (Graham et al. 1990). On the lower slopes, deep Inceptisols formed entirely in colluvium.

The degree of both profile development and weathering is minimal in Inceptisols. These marks of pedogenesis can be evaluated by the extent to which relatively easily weathered minerals have been altered. In situations where Inceptisols occupy upland positions on young geomorphic surfaces, both initial and secondary minerals are present in the soil profile. McCracken et al. (1962) concluded that feldspar, present in the C horizon but not in the solum of some Dystrudepts on steep, sloping areas of the southeastern United States, had altered to kaolinite and gibbsite in the A and B horizons. The amount of iron oxides produced by weathering of iron-bearing primary minerals is relatively low in Inceptisols. For example, Rebertus and Buol (1985b) found that the ratio of free citrate-bicarbonate-dithionite (CBD) extractable iron content to total iron was less than 0.5 in some Dystrudepts formed from mica gneiss and schist. The free-iron to total-iron ratio was lowest in those pedons judged on other criteria to be the least developed and increasing in more developed pedons to nearly 1 in some associated Hapludults. The free-iron to total-iron ratios were higher in B horizons than A horizons, indicating some iron translocation in addition to release from mineral structures by weathering.

Sand-sized biotite is frequently observed to weather to sand-sized kaolinite in Dystrudepts and associated soils (Harris et al. 1985a, 1985b; Rebertus et al. 1986; Graham et al. 1989a). Under udic soil moisture regimes, and mesic and thermic soil temperature regimes, gibbsite contents tend to be highest in the early stages of rock weathering and are often highest in the lower cambic horizon and the saprolite of Dystrudept profiles. Gibbsite, however, is apparently unstable in the acidic environment of the cambic horizon where hydroxy aluminum interlayering of vermiculite appears in the clay fraction and, perhaps, the gibbsite is resilicated to form kaolin (Rebertus et al. 1986; Graham et al. 1989b).

Clay skins are frequently observed in cambic horizons. Rebertus and Buol (1985b) found lessivage most evident in the earliest stages of Dystrudept formation and attributed the process to the rapid formation of clay from plagioclase weathering. More-developed cambic horizons, those with essentially all of the plagioclase weathered, were found to have fewer clay skins. Apparently, little clay is produced in the more mature Dystrudepts where only small amounts of plagioclase remain. Lessivage, as evidenced by clay skins, also appears to be more active in profiles where the kaolinite sand particles begin to physically break to form clay-sized kaolinite. At that stage, argillic horizons are formed and the soils classify as Hapludults.

Hillslope processes of creep, slope wash erosion, and landslides are sporadically active both in time and in space. As soil material moves, it is also actively weathering, and morphological features are being formed and destroyed. Multiple pedogenic pathways are likely as hillslope and pedogenic processes interact on steep slopes (Graham and Buol 1990).

In cryic soil temperature regimes, well-drained Inceptisols are common. They may have ochric epipedons or umbric epipedons with cambic horizons or only umbric epipedons without cambic horizons (Rieger 1983). They form mostly in loamy materials. The ochric epipedons may be formed on either felsic or mafic parent materials. Umbric epipedons, however, are formed from felsic material, whereas Mollisols form on mafic parent materials. Acid Cryepts tend to accumulate iron and aluminum near the surface, because the weathering release of these elements is most pronounced in A horizons, and downward translocation is minimal (Rieger 1983). In Arctic Cryepts, it is not uncommon for a calcic horizon to form below a cambic horizon on calcareous parent materials (Rieger 1983).

Poorly drained Inceptisols are characterized by active gleization that produces redoximorphic features. These soils are commonly located in depressional parts of the landscape where leaching may be more extensive than in other landscape positions, but the process of lessivage and thus argillic horizon formation is somewhat retarded, probably because the soils do not undergo frequent desiccation. In humid regions with felsic rocks, soils formed in landscape depressions tend to be more leached, and have exchange complexes with lower base cation content and more aluminum than soils in surrounding areas. This is especially true in regions where the surrounding better-drained soils have low base cation status and some exchangeable aluminum. In landscapes where soils are rich in base cations, the associated poorly drained Inceptisols usually have an even higher base status than the surrounding soils. This can be attributed to the enrichment of the low-lying parts of the landscape by lateral moving groundwater. In some materials saturated with brackish water, sulfides may accumulate and sulfuric horizons may be formed (Fanning and Fanning 1989). When oxidized, usually by artificial drainage, sulfuric acid is formed. These unique Inceptisols are commonly known as "cat-clays" or acid sulfate soils.

The time required to form cambic horizon features necessary to classify a soil as an Inceptisol varies widely with environmental conditions and soil processes. A number of pedogenic features can qualify subsoil as a cambic horizon. If easily altered iron-containing minerals are present, secondary oxides may form almost as quickly as rust. Inceptisols are now present in polders in the Netherlands as the result of the ripening process that occurred after the ocean was excluded by dikes (Pons and Van Der Molen 1973). Percolating water containing organic acids may quickly deplete the carbonate content of calcareous parent materials if the carbonate is finely divided, but it will take longer if the carbonate particles are coarse or if the leaching water is less acid. Reduction of iron oxide and formation of low chroma redoximorphic features can take place in red-colored soil materials in as little as one week when such material is submerged in a sealed flask and sugar is added as a carbon source to speed microbial respiration. When plant organic matter is used as the carbon source, the formation of the redoximorphic features is delayed by only several weeks (Dungan et al. 1999). The formation of redox features in the field depends on soil conditions, but iron oxide depletions have been noted to form after a single flooding event, and concentrations have been found to form within 30 years (Vepraskas 2001).

Uses of Inceptisols

Some Inceptisols have sustained agriculture for many centuries. The major river deltas of Asia have been used for flooded rice throughout recorded history. Technologies for using Inceptisols for agricultural and other uses are necessarily diverse. Some Inceptisols are formed in calcareous parent materials, only partially decalcified to form a cambic horizon, and have neutral to alkaline reaction. Others are formed in infertile, acid parent materials. Steep areas are best suited to

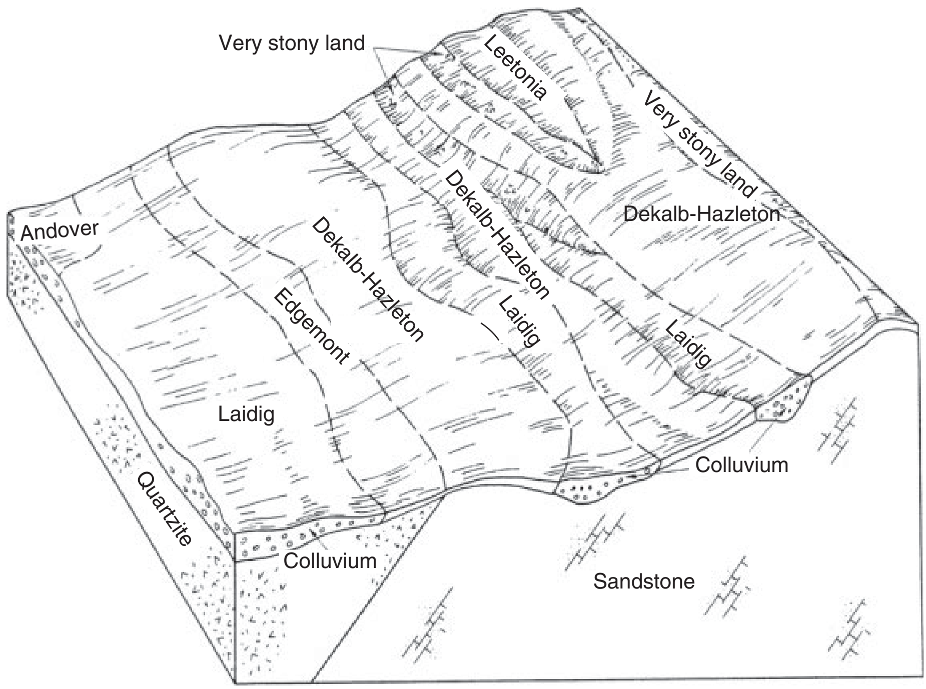
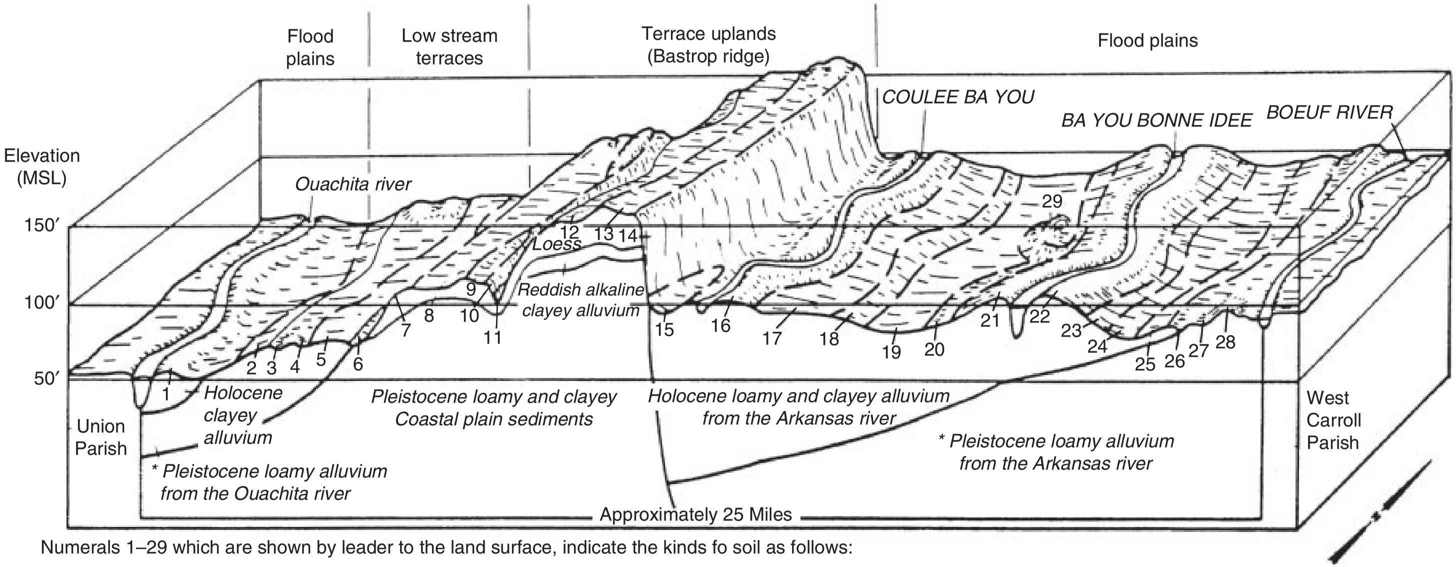


Figure 14.2. Patterns of two Inceptisols that, together with very stony land, occupy half of the area of this quite stony soilscape, three Ultisols, and one Spodosol. Dekalb (solum depth, 70 cm; depth to bedrock, 75 cm) and Hazleton (solum, 1 m thick; depth to bedrock, 150 cm) soils are loamy-skeletal, siliceous, mesic Typic Dystrudepts. Andover and Laidig soils are fine-loamy, mixed, mesic Typic Hapludults. Leetonia soils are sandy-skeletal, mixed, mesic Entic Haplorthods. The soilscape is in Franklin County, Pennsylvania. (From Long 1975)

woodland, recreation, or wildlife. Poorly drained Inceptisols can be extensively used for cultivated field crops if artificial drainage is feasible. In some landscapes, Inceptisols are more productive than associated soils of other orders. For example, associated soils with thick, rather impermeable argillic horizons may be less productive, as are associated sandy Entisols with low water-holding capacity and poor nutrient reserves.

Inceptisols of Figure 14.2 (Dekalb and Hazleton) are so steep, stony, and shallow to bedrock that they are suited only to forestry and wildlife habitat. These soils are seldom used for cultivation or home sites. Figure 14.3 shows the poorly drained Perry and Litro, somewhat poorly drained Portland, and moderately well-drained Cascilla soils. All may be used for forest and pasture. The first three support rice and corn. Portland soils are also used for production of cotton and soybeans.

Specific problems of management are associated with the Sulfaquepts and other soils with sulfuric horizons. If sulfuric horizons are created by drainage, they oxidize to form sulfuric acid, rendering the soil worthless for crop production (Bloomfield and Coulter 1973; Dost 1972).



Numerals 1–29 which are shown by leader to the land surface, indicate the kinds fo soil as follows:

1. Litro and Perry
2. Haggerty
3. Groom
4. Mollicy
5. Wrightsville and Groom
6. Lafe
7. Debute and Libuse
8. Frizzell and Guyton
9. Libuse
10. Debute
11. Guyton and Cascilla
12. Tillou and Guyton
13. Bussy
14. Udalfs
15. Allemands
16. Sterlington and Gallion
17. Mer Rouge and Gallion
18. Hebert and Portland
19. Perry
20. Hebert and Portland
21. Rilla
22. Sterlington
23. Hebert and Portland
24. Perry
25. Forestdale
26. Idee
27. Goodwill
28. Dexter
29. Yorktown

* Upper layers mixed with loess in most places.

Figure 14.3. Block diagram of a level-to-undulating terrain, which appears hilly in the drawing because vertical exaggeration is 200-fold. Twenty-six thermic soils are included, of which four are Inceptisols that occupy 33% of the area of soilscape. Three of these are Vertic Endoaquepts, namely, Perry, Portland, and Litre, which are, respectively, very fine, montmorillonitic, nonacid; very fine, mixed, nonacid; and fine, acid. The fourth (Cascilla) is a fine-silty Fluventic Dystrudept. Fifteen Alfisols include six Hapludalfs (Dexter, Frizzell, Goodwill, Gallion, Rilla, and Sterlington), four Endoaqualfs (Forestdale, Groom, Hebert, and Idee), two Fragiudalfs (Bussy, Libuse), and three others (Lafe—Natrustalf; Tillou—Glossudalf; Wrightsville—Glossaqualf). Four Ultisols include a Typic Fragiudult (Debute), an Aquic Hapludult (Mollicy), a Typic Glossaquult (Guyton), and an Aeric Endoaquult (Haggerty). Also included are Histosols (Allemands, Terric Medisaprists), Mollisols (Mer Rouge, Typic Argiudolls), and Entisols (Yorktown, Typic Fluvaquents). "Udalfs" (No. 14) is a map unit name given to areas dominated by various soils of Udalf suborders but not delineated at lower categories of *Soil Taxonomy*. The soilscape is in Morehouse Parish, Louisiana. (From Reynolds et al. 1985)

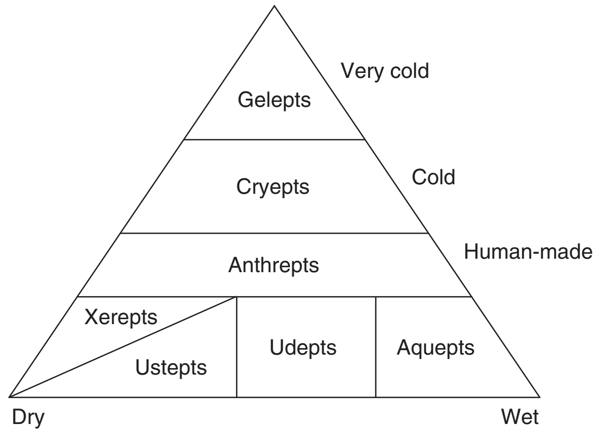


Figure 14.4. Diagram showing some relationships among suborders of the Inceptisols.

Classification of Inceptisols

Inceptisols are mineral soils that exhibit profile development sufficient to exclude them from Entisols. Most Inceptisols have a cambic horizon and thus have loamy or finer subsoil texture. Spodic, argillic, kandic, natric, and oxic horizons are not present. Inceptisols are not permitted to have andic soil properties in any layer thicker than 35 cm within the top 60 cm. Inceptisols are excluded from the Aridisols by soil moisture regime, from the Gelisols by the lack of permafrost and from the Vertisols by lack of slickensides and/or cracks that periodically open and close. Soils that lack subsoil development but have umbric, histic, or plaggen epipedons are Inceptisols. Soils with mollic epipedons are Inceptisols when base saturation at pH 7 is less than 50% in some horizon between the mollic epipedon and a depth of 180 cm or a lithic or paralithic contact if shallower. Seven suborders are separated based on soil moisture regime, soil temperature, and human influence (Figure 14.4).

Aquepts have redoximorphic features and unless artificially drained are saturated with water at some period in the year. Reaction classes are used in most Aquepts.

The suborders Udepts, Ustepts, and Xerepts are defined as having udic, ustic, and xeric soil moisture regimes, respectively.

Anthrepts have dark brown or black anthropic or plaggen epipedons. Anthrepts are formed by anthropic activity, including human-habitation and farming, mainly in Europe, but also in the United States, South America, and other parts of the world. They are of small extent.

Cryepts are Inceptisols with a cryic soil temperature regime.

Gelepts are Inceptisols with a gelic soil temperature regime.

Table 14.1 lists the great groups recognized in the Inceptisol order.

Table 14.1. Suborders and great groups in the Inceptisol order

Suborder	Great Groups
Aquepts	Sulfaquepts have a sulfuric horizon within 50 cm of the surface.
	Petraquepts have plinthite or cemented horizon that constitutes more than 50% of volume within 100 cm of the surface.
	Halaquepts have a salic horizon or 25 cm of material with ≥15% exchangeable sodium within 50 cm.
	Fragiaquepts have a fragipan within 100 cm of the surface.
	Gelaquepts have a gelic soil temperature regime.
	Cryaquepts have a cryic soil temperature regime.
	Vermaquepts have 25 cm or thicker layer within 100 cm of the surface with 25% or more (by volume) recognizable bioturbation.
	Humaquepts have a histic, melanic, mollic, or umbric epipedon.
	Epiaquepts have episaturation.
	Endoaquepts—other Aquepts (have endosaturation).
Anthrepts	Plagganthrepts have a plaggen epipedon.
	Haplanthrepts—other Anthrepts.
Gelepts	Humigelepts have an umbric or mollic epipedon.
	Dystrogelepts have <50% base saturation (NH_4Oac) within 50 cm of surface.
	Haplogelepts—other Gelepts.
Cryepts	Humicryepts have an umbric or mollic epipedon.
	Calcicryepts have a calcic or petrocalcic horizon within 100 cm of surface.
	Dystrocryepts have no free carbonates within 200 cm of the surface and <60% base saturation (NH_4Oac) in most of the 25–75 cm depth zone.
	Haplocryepts—other Cryepts.
Ustepts	Durustepts have a duripan within 100 cm of the surface.
	Calciustepts have a calcic horizon within 100 cm, or petrocalcic horizon within 150 cm, of the surface and either are calcareous or have a texture of loamy fine sand or coarser in all parts above the calcic or petrocalcic horizon after mixing the top 18 cm.
	Humustepts have an umbric or mollic epipedon.
	Dystrustepts have no free carbonates within 200 cm of surface and <60% base saturation (NH_4OAc) in all of the 25-75 cm depth zone.
	Haplustepts—other Ustepts (have base saturation ≥60%).
Xerepts	Durixerepts have a duripan within 100 cm of the surface.
	Fragixerepts have a fragipan within 100 cm of the surface.
	Humixerepts have an umbric or mollic epipedon.
	Calcixerepts have a calcic horizon within 100 cm, or petrocalcic horizon within 150 cm, of the surface and are calcareous in all parts above those hoizons after mixing the top 18 cm.
	Dystroxerepts have no free carbonates within 200 cm of surface and <60% base saturation (NH_4OAc) in all of the 25–75 cm depth. zone.
	Haploxerepts—other Xerepts (have base saturation ≥60%).
Udepts	Sulfudepts have a sulfuric horizon within 50 cm of the surface.
	Durudepts have a duripan or other cemented horizon within 100 cm of the surface.
	Fragiudepts have a fragipan within 100 cm of the surface.
	Humudepts have an umbric or mollic epipedon.
	Eutrudepts have free carbonates or base saturation ≥60% in some horizon between 25 and 75 cm.
	Dystrudepts—other Udepts (have base saturation <60%).

Perspective

Perhaps because of the derivation of their names from the Latin *inceptum,* beginning, Inceptisols are often portrayed as soils on their way to developing into some other kind of soil. In many environments, such a transformation probably will not take place until one or more of the factors controlling pedogenic processes changes.

As a pedogenic concept, Inceptisols include soils that have some subsoil development, the cambic horizon, but lack features thought to represent mature soil formation. This conceptual framework led to the inclusion of diverse pedons, although a large part of the original diversity was eliminated when Andepts were reclassified into the Andisol order. Further reduction in diversity followed the adoption of the Gelisol order, which removed soils with permafrost from the Inceptisol order. The remaining Inceptisols can then be characterized as mineral soils, rich in weatherable minerals with minimal subsoil development. They are of medium to fine texture, not sandy, and have sufficient natural moisture for one or more crops per year where temperatures permit.

Relatively few pedogenic studies have been directed to Inceptisols. They offer future students unique opportunities to study the weathering of easily weathered minerals and better understand the early stages of many pedogenic processes.

Mollisols: Grassland Soils of Steppes and Prairies

15

Mollisols (from Latin *mollis*, soft) are characterized by having a deep, dark, friable and relatively fertile surface horizon (or horizons) known as a *mollic epipedon* (Figure 15.1; Figure 15.2). The vast majority of Mollisols are formed under grassland vegetation, where the annual proliferation of fine roots contributes to a relatively high organic carbon content. Other Mollisols may include soils of poorly drained lowland hardwood forests and some well-drained forested soils, often with significant understory vegetation. In addition to a mollic epipedon, Mollisols are characterized by relatively high base status to a considerable depth. Accordingly, these soils possess a high level of native fertility that has been widely exploited for agricultural production, often with minimal inputs of lime and fertilizers. There is also considerable biological activity associated with Mollisols, with earthworms, rodents, and various insects typically playing an important role in the formation of these soils.

Setting

Mollisols occupy approximately 7% or slightly more than 9,128,000 km^2 of the ice-free global land to area (as shown in Table 20.3) and occur most commonly in the temperate grasslands of the middle latitude. These ecosystems occupy as much as 15,100,000 km^2 of the global land area (Schlesinger 1997) and are extensive in North America, South America, Asia, and Europe. Ecologically and climatically, midlatitude grasslands represent the broad expanses between drier desert and moister forest communities, and include both the short-grass steppe and the tall-grass prairie. Although Mollisols are the dominant soils of these ecosystems, extensive areas of Aridisols and Entisols can be found in the drier steppe. Alfisols are common in the moister regions where the tall-grass prairie gives way to forest.

The short-grass grassland, or steppe, often resembles a pastured meadow extending monotonously to the horizon where grasses typically stand 15- to 30-cm high. Only in unusually wet years do patches of taller grasses develop enough to give the vegetative cover an uneven appearance. Sagebrush (*Artemisia* spp.) is a major component of the drier steppe regions of the western United States and, where dominant, gives the landscape a shrubby appearance (Fosberg 1965). Blue grama grass (*Bouteloua gracillis*) is common on drier Ustolls (Thorp 1948), and small soapweed (*Yucca glauca*) is prominent in places. Its roots spread as much as 2 m vertically and

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Figure 15.1. Profile photo of a Mollisol (Cryoll) from Montana showing black soil colors in the top 20 cm. Average organic matter content in the top 20 cm is approximately 6.8%. Scale is in decimeters (C) and feet (F).

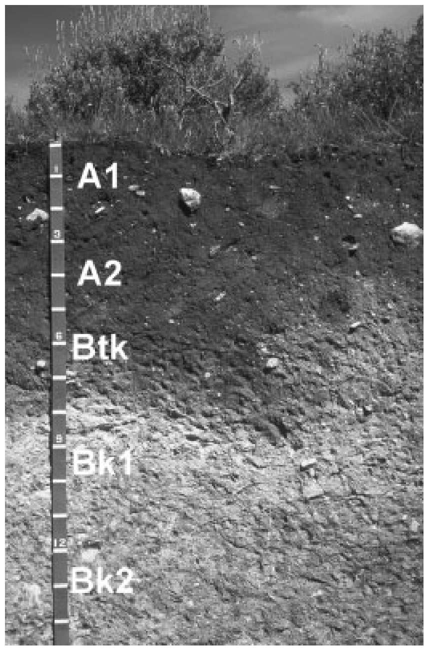

Figure 15.2. Pachic Argicryoll from Lemhi County, Idaho. Soil has formed in glacial drift and has a thick (pachic) mollic epipedon. Mean annual precipitation is ~430 mm; native grasses and sagebrush are the dominant vegetation. The upper right-hand side of the profile has been extensively mixed by badgers. For color detail, please see color plate section.

10 m laterally. Western wheatgrass (*Pascopyrum smithii*) and buffalo grass (*Bouteloua dactyloides*) are found on Ustolls. Buffalo-grass sod was used by pioneers for building houses. *Stipa* is the main genus in the western Russian steppes. The mean annual temperatures and precipitation of 2°C and 200 mm at Urga, Mongolia, 14°C and 250 mm at Quetta, Pakistan, and 4°C and 360 mm at Williston, North Dakota, represent climatic conditions on middle-latitude steppes, respectively (Finch et al. 1957). In the Great Plains of North America, mean annual precipitation may be as low as 250 mm at the steppe-desert boundary along the western margin (Anderson 1987).

The tall-grass prairie is grassland of relatively luxurious growth of vegetation that stands 1- to 3-m high at maturity. The natural stands of the Argentine Pampas were so tall that a person riding on horseback could disappear from sight. Big and little bluestem grasses (*Andropogon gerardii* and *Schizachyrium scoparium*) are among the tall grasses found on Udolls of the Great Plains. Tall-grass prairies develop under relatively moist conditions: 8°C and 810 mm in eastern Iowa and 16°C and 760 mm in central Oklahoma. On the eastern Great Plains, the tall grass prairie is replaced by forest where mean annual precipitation exceeds 750–1,000 mm (Anderson 1987).

Borchert's analysis (1950) of the climate of the prairie triangle or 'peninsula' of the Great Plains of North America lists the essential ingredients of the climates of middle latitude grasslands. These features occur in different proportions over these lands: (1) severe, dry winters with much wind and relatively slight accumulations of snow; (2) relatively moist springs in most years; and (3) droughty summers with some thunderstorms and tornadoes. It is also important to recognize that some Mollisols exist in tropical areas of the world, although on a global basis, only accounting for approximately 4% of all Mollisols. Many tropical Mollisols have formed on calcareous parent materials most commonly under udic and ustic soil moisture regimes, and have organic carbon contents that rival or exceed those of their temperate counterparts.

The boundaries between desert, steppe, tall-grass prairie, and forest are often irregular and have not remained stationary over time. Complexities in the geographical distribution of parent materials, topography, climate change, and fire history are among the reasons for this. On the borders of drier zones, coarse-textured soils allow greater infiltration and deeper penetration of sporadic rains, thereby favoring extension of grass into drier regions. Similarly, on the borders of more humid zones, coarse-textured soils favor forest growth in the prairie lands, as in the case of the cross-timbers of Texas (Thorp 1948).

There is considerable evidence that grassland boundaries have migrated back and forth over time. As a result, it is likely that many older Mollisols have developed under more than one climatic regime and plant community. Dry-to-moist climate changes during the Holocene have resulted in forest encroachment into prairie (Fenton 1983). Periods of warmer and drier conditions such as the altithermal period have also allowed prairie expansion (Ashworth and Brophy 1972). Curtis (1959) refers to the altithermal period as the great period of prairie expansion. Significant portions of the Lake Michigan basin were occupied by soils, with the lake (called Lake Chippewa) being small and standing at 75 m (230 ft) above sea level.

Table 15.1. Selected properties of Mollisols. Data are for the Holdrege series (Pedon no. 82P0772), and the Palouse series (Pedon no. 86P0071) are from the USDA-NRCS National Cooperative Soil Survey Soil Characterization Database (Soil Survey Staff 2010b)

Horizon	Depth cm	Moist color	Structure[a]	Texture	Bulk density g cm^{-3}	pH H_2O	Organic C %	Exchangeable bases[b] Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	Acidity	CEC pH7	CaCO$_3$ %	Base saturation[c] %
								cmol (+) kg^{-1}							
Holdrege series (Typic Argiustoll)—Nebraska															
A1	0–19	10YR 2/2	2 f gr	silt loam	1.26	5.9	2.23	12.9	3.9	trace	1.6	6.2	20.1		75
A2	19–33	10YR 3/2	2 m gr	silty clay loam	1.27	6.7	1.30	15.7[d]	5.6	0.1	1.9	3.4	23.0		87
Bt1	33–44	10YR 3/3	2 f sbk	silty clay loam	1.29	7.0	0.85	16.1[d]	6.3	0.1	1.9	2.7	24.4	—[e]	90
Bt2	44–67	10YR 5/2	2 m sbk	silt loam	1.32	7.4	0.40	15.0[d]	6.2	0.1	2.0	2.4	22.6	—	91
Bt3	67–92	10YR 4/3	1 m sbk	silt loam	1.35	7.8	0.24	14.9[d]	6.5	0.3	2.2	1.1	21.7	—	96
BC	92–142	10YR 5/3	1 m sbk	silt loam	1.34	8.3	0.14	29.6[d]	6.6	1.0	2.2		19.7	2	100
Palouse series (Pachic Haploxeroll)—Washington															
A1	0–10	10YR 2/1	3 f gr	silt loam	0.94	6.6	3.10	22.4[d]	4.1	0.2	1.8	4.9	26.4		85
A2	10–25	10YR 2/1	3 m gr	silt loam	1.18	6.8	3.13	19.6	4.2	0.1	1.2	4.5	26.2		85
A3	25–48	10YR 2/2	2 m sbk	silt loam	1.18	6.3	1.97	17.9	4.0	0.1	0.5	5.9	24.8		79
AB	48–74	10YR 3/2	2 m pr	silt loam	1.16	6.3	1.07	17.4	4.4	0.1	0.3	4.6	23.0		83
BA	74–79	10YR 3/2	2 m sbk	silt loam	1.32	6.9	0.56	17.1[d]	5.0	0.2	0.2	3.2	22.5		88
Bw1	79–107	10YR 4/3	2 m sbk	silt loam	1.31	7.0	0.35	16.7[d]	4.7	0.4	0.1	2.8	21.8	—	89
Bw2	107–140	10YR 4/3	1 m sbk	silt loam	1.36	8.2	0.27	19.7[d]	4.8	1.0	0.1	1.2	23.5	—	96

[a] 1 = weak; 2 = moderate; 3 = strong; f = fine; m = medium; gr = granular; sbk = subangular blocky; pr = prismatic.
[b] extracted with ammonium acetate.
[c] by sum of cations.
[d] may include Ca from calcium carbonate or gypsum.
[e] none detected.

Fire, both natural and anthropogenic, is also an important agent in many grassland-forest ecotones. At the edges of the grasslands, such as with the boundary of deciduous forest in Wisconsin, extensions of prairies by fire have formed preferentially on topography over which fire moves easily, namely ridge tops and some windward slopes. Advance of aspen forest into prairie regions has been observed in Canada following a reduction in prairie fires that coincided with settlement in the early 1900s (Bird 1961, cited in Anderson 1987).

Mollisols occur on deposits and landscapes with a wide range of ages. Many are formed in Holocene-age deposits associated with glaciation. Others, especially those that have argillic horizons, occupy older deposits and landscapes (probably late Pleistocene) that have experienced variation in climate and vegetation. These Mollisols are clearly polygenetic and probably were under forest vegetation during glacial periods (Fenton 1983). Development of polygenetic Mollisols in the Palouse prairie region of the Pacific Northwest United States may span a period of approximately 40,000 years (McDaniel and Hipple 2010). Some Russian soil scientists suggest that the post-glacial-age Mollisols evolved during a changing climate. A poorly drained condition gave way to better-drained conditions as the climate became warmer and drier. The soils became alkaline, and then dealkalized (solodized) and were left with the present carbonate-rich condition.

Pedogenic Processes

Melanization, the process of darkening of the soil by addition and decomposition of organic matter, is the dominant process in Mollisols. It is the process by which the mollic epipedon forms and dark soil colors extend down into the profile. Soil color, structure, and organic carbon data presented in Table 15.1 reflect the strong influence of melanization in the A horizons of two Mollisols formed under grassland vegetation. Melanization is actually a bundle of several more-specific processes including extension of roots of prairie vegetation into the soil profile; microbial decomposition of organic materials in the soil, producing some relatively stable, dark compounds (humification); and reworking of the soil and organic materials by earthworms, ants, cicada nymphs, and rodents (bioturbation) (Hole and Nielsen 1970). In Mollisols, melanization is driven primarily by the incorporation of organic matter directly into the mineral soil.

It may be somewhat surprising that melanization and the accompanying accumulation of organic carbon is such a dominant process in temperate grassland soils, given the relatively low net primary production of these ecosystems (Schlesinger 1994). However, despite the relatively low net primary production, the net annual addition of carbon to Mollisols typically exceeds that for soils of tropical and temperate forests (Bolin et al. 1979). This has been attributed to the high proportion of material derived from roots coupled with relatively low rates of decomposition (Oades 1989). Roots are as much as 80% of the total biomass in many grasslands (Lauenroth and Whitman 1977; Fenton 1983). Thorp (1948) estimated that annual additions of raw organic matter to Ustoll soil profiles ranged from 590 to 1030 kg dry weight ha^{-1} (520 to 900 lb $acre^{-1}$), mostly by in situ root death, and as much as 1250 kg ha^{-1} organic matter may be added annually to Udolls of the tall-grass prairies. Both the depth of rooting and quantity of roots have been strongly correlated with the mollic epipedon thickness (Cannon and Nielsen 1984).

Numerous studies have documented the rapid accumulation of organic carbon in Mollisols. Schafer et al. (1980) found that the amount of organic carbon accumulation in the 0-to-10-cm depth of 50-year-old soils formed in mine spoils was similar to that of nearby reference soils (Ustolls). Ulery et al. (1995) showed that organic carbon content more than doubled in developing Mollisols after just 41 years.

Organic materials undergo significant change once added to the soil. The stable humus formed during melanization is a combination of the less palatable parts of the original organic matter, plus complex organic compounds synthesized by soil microorganisms (Oades 1989). Many of these resistant organic compounds are polymers of phenolic and aromatic functional groups (Martin and Haider 1971). The association between clays and aromatic humic substances in the Ca-rich environment (Table 15.1) afforded by Mollisols produces aggregates that are resistant to physical disintegration and further biological change. This stability is reflected in the average age of organic carbon in Mollisols, as assessed by radiocarbon dating techniques. Soil organic carbon of Mollisols (and Histosols) is older than that of other soil types (Oades 1989). Average ages of stable organic carbon in Mollisols from the Great Plains range from several 100s of years up to 3,000 years (Hseih 1992).

Wet conditions lead to increased production of plant biomass, decreased turnover, and subsequently greater accumulation of soil organic matter. Lowering of soil redox status curtails aerobic decomposition, thereby reducing the efficiency and rate of decomposition. In addition, the higher heat capacity of wet soils results in lower maximum temperatures, which also reduces decomposition rates. These conditions are responsible for the occurrence of Mollisols in the most poorly drained positions of Alfisol-dominated landscapes of the Midwest United States (Brown and Thorp 1942). In Mollisols of Iowa, increasing periods of saturation and reduction are related to higher soil organic carbon contents (Khan and Fenton 1994).

Colder temperatures are also responsible for the high organic carbon contents of some Mollisols (McDaniel and Munn 1985). Microbial activity is reduced in many Mollisols having frigid, cryic, and gelic soil temperature regimes, and this contributes to the relatively large quantities of organic carbon contained in these soils. This mechanism also contributes to the formation of some forested Mollisols, such as those that occur in Alaska under spruce, birch, and aspen (Soil Survey Staff 1999).

Where carbonates are present in the parent materials, translocation of carbonates is a common pedogenic process in Mollisols. The depth at which $CaCO_3$ has been accumulated is a good general indicator of mean annual precipitation (Jenny 1941). In the Mollisols receiving higher quantities of rainfall (Udolls), all $CaCO_3$ is usually leached from the soil profile. In contrast, Mollisols of drier environments typically contain $CaCO_3$ at or close to the soil surface, with the pattern of increasing depth to

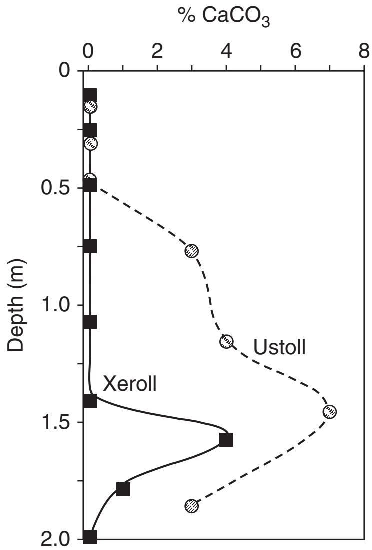

Figure 15.3. Depth distribution of $CaCO_3$ in Mollisols formed in calcareous loess under approximately 540 mm of mean annual precipitation. Xeroll (86P0071) is from eastern Washington and receives the majority of precipitation during the winter months. Ustoll (40A1862) is from western Kansas and receives most of the precipitation during the growing season. Data are from the USDA-NRCS National Cooperative Soil Survey Characterization database (Soil Survey Staff 2010b).

$CaCO_3$ with increasing annual precipitation (Munn et al. 1978). In addition, the seasonal distribution of precipitation exerts a strong control over carbonate translocation. For a given amount of mean annual precipitation, carbonates are leached to a greater depth in soils that receive much of their rainfall during the winter months when evapotranspiration is low (Figure 15.3). This pattern is characteristic of many Mollisols of xeric soil moisture regimes and is indicative of the efficiency of leaching that can occur under this regime with relatively low total annual precipitation.

Translocation of clays (lessivage) may occur in some Mollisols, but the expression of this process is extremely variable. Colloidal clays tend to exist in close association with humic substances in Mollisols, forming organo-clay complexes. It has been suggested that because of the presence of these complexes and the rapid absorption of water by plant roots, lessivage occurs slowly in Mollisols. And, as discussed below, the activities of ants and earthworms may inhibit the formation of argillic horizons as well.

Argillic horizons are typically found in Mollisols that occupy older, more stable landscapes. Significant lessivage can only occur after $CaCO_3$ has been leached from a soil horizon and clays are able to disperse (Fanning and Fanning 1989). Because $CaCO_3$ leaching increases with increasing precipitation, significant translocation of clay tends to be more common in Mollisols of higher-rainfall areas. Argillic horizons that occur in semiarid Mollisols are typically found immediately above Bk horizons.

Eluviation and illuviation of organo-clay complexes results in the surfaces voids between B horizon peds becoming coated with dark cutans (organo-argillans). This process contributes to melanization (Hole and Nielsen 1970). The organo-argillans on B-horizon ped surfaces indicate that blocky and prismatic soil structures are prerequisite to the development of the coatings. The coatings in turn may favor the maintenance of the structure (Wittmuss and Mazurak 1958).

In general, relatively little mineral weathering has occurred in Mollisols of temperate regions. This can be attributed to relatively dry, cool climatic conditions and the fact that many Mollisols are formed in geologically young surficial deposits. As a result, minerals tend to be inherited from the parent material, and clay mineralogy typically includes clay mica (illite), vermiculite, and smectite (Bell and McDaniel 2000). Kaolinitic and halloysitic mineralogy has been observed in Mollisols of the tropics, suggesting that considerable weathering may occur in these soils or the parent material was almost devoid of weatherable minerals prior to soil formation.

Activity of earthworms is considerable in some Mollisols. Baxter and Hole (1967) noted earthworms and their casts even in active ant mounds. Buntley and Papendick (1960) used the term Vermisol for soils (now Vermustolls) that had been thoroughly worked to a depth of 60 cm or so by earthworms. They also found evidence that such extensive earthworm activity may inhibit argillic horizon formation. These soils showed no textural change in the solum and had B horizons with granular rather than prismatic structure, a dispersed Bk horizon, and an unusually thick A horizon. It is possible that earthworms bring some $CaCO_3$ to the surface in Ustolls, especially the Vermustolls. In California, earthworm activity in developing Mollisols preferentially moves clay-size material to the surface in stable worm casts (Graham and Wood 1991). This counteracts any clay illuviation that has taken place and inhibits further eluviation of clays, making argillic horizon formation unlikely or extremely slow in these soils.

Observations and calculations by Baxter and Hole (1967) indicate ant activity can result in the translocation of clay from the B horizon to the A horizon. The common prairie ant, *Formica cinerea,* builds mounds as much as a foot high and at least as wide (Figure 15.4). Clay content in the mounds is equivalent to that in the B horizon. In addition, argillans characteristic of B and C horizons have been reported from thin sections of ant-mound material. Contents of available phosphorus and potassium are extremely high in the mounds, possibly in part because the mounds are largely composed of yellowish B horizon material, and in part because the ants concentrate organic materials from aphids, vegetation, and from their own bodies in the mounds. This process by which ants translocate clay from B to A horizons can account for the presence of A and B horizons with nearly equal clay contents in many medium-textured, well-drained Mollisols of the upper Midwest United States (Baxter and Hole 1967).

Some Xerolls in the western United States have B horizons that consist almost entirely of cicada burrows; often these cylindrical burrows are cemented with carbonate and silica and are found at depths up to 1–2 m (Hugie and Passey 1963). Because active cicadas in the western United States generally burrow within the 0.5 m of soils, these deep burrows are interpreted as evidence of buried soils in aggrading landscapes (O’Geen and Busacca 2001).

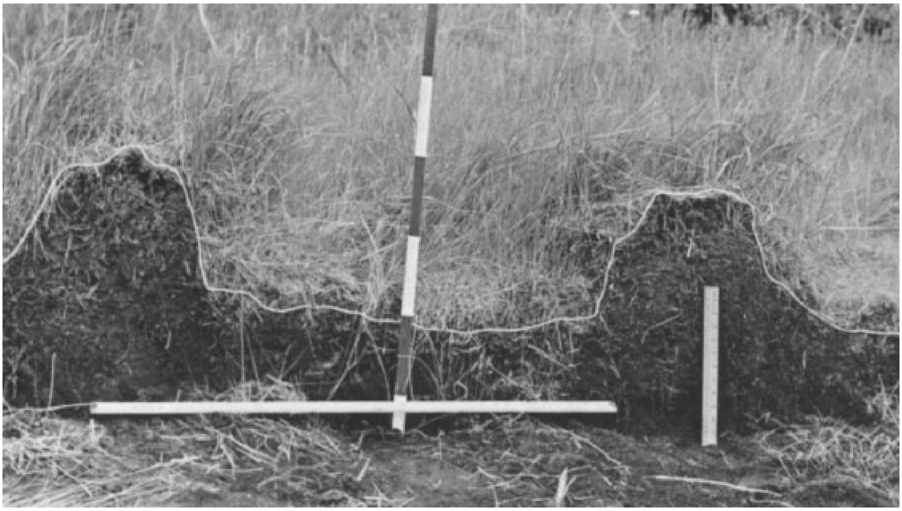

Figure 15.4. Cross section through two mounds of the common western mound-building ant. (After Baxter and Hole 1967)

Rodents are also active in many Mollisols and may rework considerable quantities of soil material (see Figure 15.2). It has been estimated that rodents annually bring to the surface as much as 20 to 40 tons of subsoil material (air-dry weight) per acre (Thorp 1949; Matelski 1959). Laycock (1958) noted that gophers in Wyoming burrow in snow and may fill the tunnels in the snow with soil, which is then left on the surface of the ground after the snow melts. Curtis (1959) estimated that the upper 60 cm of a Hapludoll was turned over once each century by the combined activity of ants, worms, and rodents.

Cumulization, or the addition of mineral material to the soil surface, can be an important process in many Mollisols. Erosion occurring along hillslopes can result in the overthickening of the mollic epipedon of soils occupying depositional landscape positions (Figure 15.5). This process can be important at the field scale, where colluvium and hillslope alluvium accumulate on toeslopes of cultivated fields, and on a broader, watershed scale where colluvial toeslopes merge with alluvial floodplains. Overthickening of the mollic epipedon by cumulization is recognized by Cumulic subgroups in several of the Mollisol great groups.

Mollisols of the short-grass steppes of North America have small increments of loess added to the A horizon annually. These new materials are 'pedogeneticized' as they are added (Schaetzl and Anderson 2005). As a result, the A horizons may continually grow upward, and horizons of calcium carbonate or clay accumulation also shift upward. This may produce a younger B horizon above an older B horizon, and in some cases, overlapping or welded B horizons. McDonald and Busacca (1990) suggest that significant horizon development in Mollisols of aggrading loessial landscapes has occurred only when pedogenic processes have been dominant over

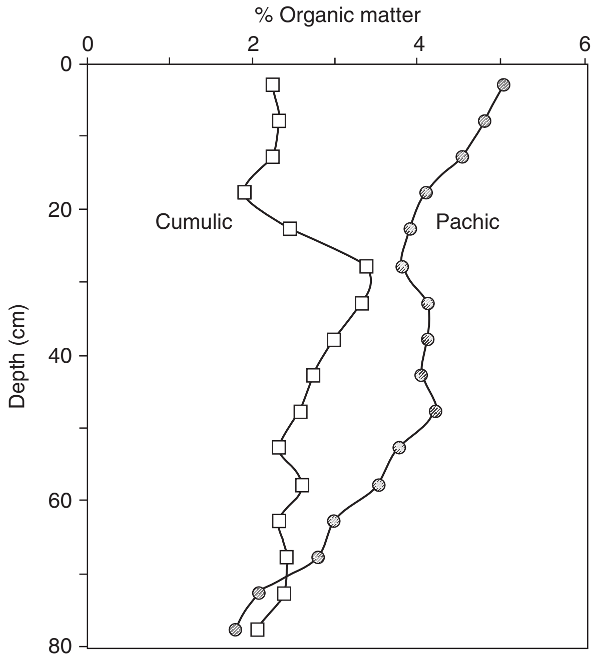

Figure 15.5. Organic matter distribution with depth in thick mollic epipedons of Mollisols from the Palouse prairie region of the Pacific Northwest U.S. Pachic soil is on a stable upland position and exhibits a relatively uniform decrease in organic mater with depth. Cumulic soil is on a footslope position where cumulization has resulted in the irregular decrease with depth. (Data courtesy of University of Idaho Pedology Laboratory)

depositional processes, such as during periods when loess deposition rates have been low. When new surface accumulation rates are rapid, pedogenesis may be unable to keep pace. In such a scenario, referred to as *retardant upbuilding* (Schaetzl and Anderson 2005), existing horizons are buried and effectively isolated beyond the range of pedogenic processes. An example of a polygenetic Mollisol formed in this manner is shown in Figure 15.6. This soil has developed in two distinct loess units. A mollic epipedon and cambic horizon have formed in the younger loess; these overlie dense, buried Bt horizons that formed in the older loess unit. An albic E horizon has formed in what was presumably the A horizon of the older buried soil and the base of the L1 loess unit; redoximorphic processes associated with episaturation are responsible for the E horizon genesis (Kemp et al. 1998).

Uses of Mollisols

To a notable extent, humans utilize Mollisols for food production. The base status (base saturation percentage) of these soils is high. Early farmers were quick to realize that these soils would be productive once the tough sod could be broken by the plow. Aboveground biomass clearing could easily be accomplished by fire. Mollisols were first farmed with little or no additions of fertilizer. But for modern high yields, significant quantities of complete fertilizer are required. In the United States today, cultivated crops have largely replaced the native grasses. The Udolls and drained Aquolls produce a large fraction of the corn and soybean crop. The Xerolls, Ustolls,

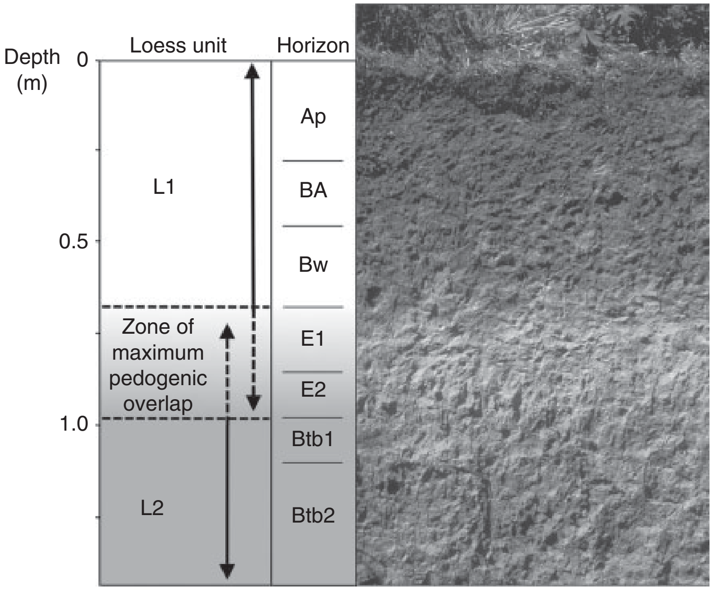

Figure 15.6. Loess stratigraphy, horizonation, and polygenetic Mollisol of an aggrading loessial landscape. Pedogenesis in the younger L1 unit (~0–15 ka) has formed a mollic epipedon and a cambic horizon, which overlie an argillic horizon formed in the L2 (gray shading) unit (~15–70 ka). Albic E horizons have developed in the lower L1 and upper L2 units as a result of episaturation. The soil is an Oxyaquic Argixeroll (Southwick series). (Adapted from Kemp et al. 1998; McDaniel and Hipple 2010)

and some Cryolls produce much of the U.S. wheat crop and significant crops of alfalfa for hay. The drier Xerolls and Ustolls generally require supplemental irrigation for crops other than grains, but can nevertheless be very productive soils. Xerolls of the Palouse region of eastern Washington and northern Idaho are among the most productive wheat-producing dryland soils in the world.

As testament to their productivity, virgin areas of Mollisols are exceedingly rare. For example, it is estimated that only about 0.1% of the Palouse prairie region of Washington, Idaho, and Oregon remains undisturbed by agricultural and urban activities (Noss et al. 1995). Relatively undisturbed areas of Mollisols are available for study only in corners of cemeteries and along little-used railroad tracks and rough lands such as the Flint Hills of Kansas. Russia has preserved several areas for scientific study.

The extensive use of Mollisols for agricultural production has had significant impacts on soil properties. Cultivation of Mollisols in the Great Plains of North America has reduced organic carbon contents by as much as 35% in approximately 70 years (Tiessen et al. 1982). In Illinois, cultivated and artificially drained Mollisols have soil C and N concentrations typically 30 to 50% less than virgin prairie soils; however, organic matter pools in these soils have apparently been stable since the 1950s when practices utilizing synthetic fertilizers were adopted (David et al. 2009). The loss of carbon is attributable to several factors related to agricultural production, including replacement of native grasses and their root systems by annual crops wherein much of the biomass is harvested and removed, increased runoff and erosion, artificial drainage, and increased decomposition rate of soil organic matter as bare soil surfaces acquire higher diurnal temperatures. In the most extreme cases of accelerated erosion, much or all of the original mollic epipedon has been lost from these soils. This has resulted in mollic epipedons being transformed into ochric epipedons, thereby excluding these soils from the Mollisol order and placing them in the Alfisol or Inceptisol orders (Fenton 1983).

The relationship between productivity and organic carbon content of Mollisols is one that deserves closer analysis. It is a commonly held belief that because Mollisols possess high levels of soil organic carbon, they are therefore very fertile and productive soils. This view is usually predicated on the fact that organic carbon content can be positively correlated to a number of soil properties important to plant growth (Brady and Weil 1999). However, it is important to recognize that the accumulation of organic carbon represents the long-term equilibrium between production and decomposition of organic residues in a soil, often over millennia. Therefore, the high levels of organic carbon observed in a Mollisol are perhaps better interpreted as an integrated record of past site productivity. As such, the expression of a mollic epipedon reflects numerous site variables such as length of growing season, soil moisture availability, high base saturation percentage, and soil temperature, to name only a few. It is these variables rather than organic matter content *per se*, that ultimately determine the productivity of a site. So while Mollisols are typically thought of as fertile and productive soils, these qualities are largely the result of high-base status parent materials and other favorable site characteristics. Accordingly, the mollic epipedon with its high organic carbon content should be regarded as a fossil record of productivity rather than the cause of it.

Classification of Mollisols

In general, Mollisols are thought of as those soils that have mollic epipedons. While it is true that all Mollisols have mollic epipedons, the presence of a mollic epipedon does not automatically qualify a soil as a Mollisol. Mollisols must have a base saturation of 50% or more in all horizons to a depth of 180 cm or a lithic or paralithic contact if shallower (Soil Survey Staff 2010a). The 50% base saturation requirement used to define Mollisols and mollic epipedons is determined by the NH_4OAc method (CEC_7), making it roughly equivalent to the 35% base saturation by sum of cations ($CEC_{8.2}$) that is used to differentiate Alfisols and Ultisols. (See Chapter 2.) Mollic epipedons are present in many Vertisols, in which case the plastic, shrink-swell nature of the clay is a more significant soil property than the mollic epipedon. Also, mollic epipedons are found in the Inceptisols with acid cambic horizons that more significantly influence the profile than does the mollic epipedon that in some cases may have been formed by the common practice of lime applications (Smith 1965). A few Alfisols also have mollic epipedons where nutrient cycling has extensively

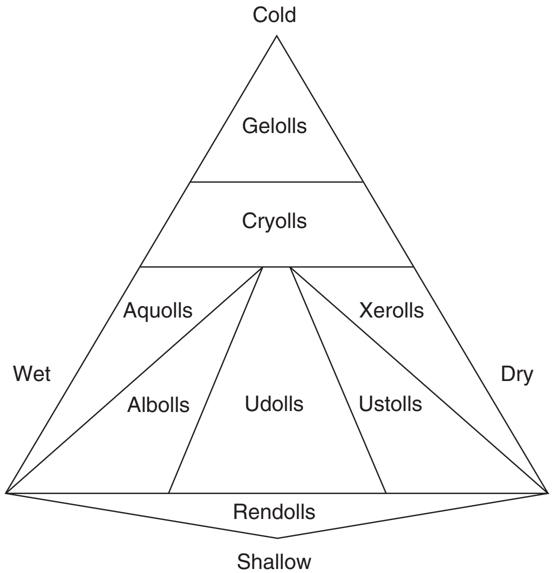

Figure 15.7. Diagram showing relationships among suborders of Mollisols.

removed bases from the subsoil and concentrated them in the epipedon, creating a base saturation percentage less than 50 in the upper part of the argillic horizon (see Figure 15.2). It is also noted that epipedons that are made to meet the mollic criteria by the common practice of agricultural liming are excluded from criteria when placing a soil in the Mollisol order. Mollisols may or may not have duripans, albic, argillic, calcic, petrocalcic, gypsic, cambic, and natric horizons. Mollisols have received considerable attention in *Soil Taxonomy*, where they are classified into the largest number of subgroups of any order.

Mollisols are subdivided into eight suborders (Figure 15.7): Albolls, Aquolls, Rendolls, Gelolls, Cryolls, Xerolls, Ustolls, and Udolls (listed in the order in which they key out in *Soil Taxonomy* and appear in Table 15.2).

Albolls are the Mollisols with an albic horizon, aquic conditions for some time in most years, and redox concentrations within 100 cm of the mineral soil surface. Below the albic horizon there is an argillic or natric horizon. These soils have formed on broad, nearly level interfluve ridge tops or closed depressions. The eluviation-illuviation and leaching processes are enhanced by the larger amounts of water moving through the profile relative to adjoining areas. Significant reduction of iron during periods of water saturation is an important process for the formation of the albic horizon. These soils were classed as Planosols or claypan soils in earlier American soil science literature. The Albolls are subdivided into two great groups: Argialbolls and Natralbolls. The Albolls are least extensive of the Mollisols, occupying only 0.01% of the earth's land area. (See Table 20.3.)

Aquolls are the Mollisols that have aquic conditions and soil properties associated with wetness, including a histic epipedon overlying the mollic epipedon, accumulation of calcium carbonate or exchangeable Na near the soil surface, or

Table 15.2. Suborders and Great Groups in the Mollisols Order

Suborder	Great Group
Albolls	Natralbolls have a natric horizon.
	Argialbolls—other Albolls.
Aquolls	Cryaquolls have a cryic temperature regime.
	Duraquolls have the upper boundary of a duripan within 100 cm.
	Natraquolls have a natric horizon.
	Calciaquolls have the upper boundary of a calcic or gypsic horizon within 40 cm and do not have an argillic horizon.
	Argiaquolls have an argillic horizon.
	Epiaquolls have episaturation.
	Endoaquolls—other Aquolls (have endosaturation).
Rendolls	Cryrendolls have a cryic soil temperature regime.
	Haprendolls—other Rendolls.
Gelolls	Haplogelolls have a gelic soil temperature regime.
Cryolls	Duricryolls have the upper boundary of a duripan within 100 cm.
	Natricryolls have a natric horizon.
	Palecryolls have the upper boundary of an argillic horizon at least 60 cm below the soil surface.
	Argicryolls have an argillic horizon.
	Calcicryolls have the upper boundary of a calcic or petrocalcic horizon within 100 cm and are calcareous in all overlying horizons or have a texture coarser than loamy fine sand.
	Haplocryolls—other Cryolls.
Xerolls	Durixerolls have a duripan within 100 cm of the soil surface.
	Natrixerolls have a natric horizon.
	Palexerolls have the upper boundary of a petrocalcic horizon within 150 cm or an argillic horizon, with some 7.5YR or redder colors, within which the maximum clay content does not decrease by more than 20% within 150 cm, or with at least 35% clay and an abrupt clay content increase of at least 15%.
	Calcixerolls have the upper boundary of a calcic or gypsic horizon within 150 cm and are calcareous in all overlying horizons or have a texture coarser than loamy fine sand.
	Argixerolls have an argillic horizon.
	Haploxerolls—other Xerolls.
Ustolls	Durustolls have the upper boundary of a duripan within 100 cm.
	Natrustolls have a natric horizon.
	Calciustolls have the upper boundary of a calcic or gypsic horizon within 150 cm and are calcareous in all overlying horizons or have a texture coarser than loamy fine sand.
	Paleustolls have the upper boundary of a petrocalcic horizon within 150 cm or an argillic horizon, with some 7.5YR or redder colors, within which the maximum clay content does not decrease by more than 20% within 150 cm, or with at least 35% clay and an abrupt clay content increase of at least 15% (absolute).
	Argiustolls have an argillic horizon.
	Vermustolls have a mollic epipedon that below 18 cm contains 50% or more (by volume) wormholes, worm casts, or filled animal burrows and rests on either a lithic contact or an underlying layer with 25% or more of similar features.
	Haplustolls—other Ustolls.

Table 15.2. *Concluded.*

Suborder	Great Group
Udolls	Natrudolls have a natric horizon.
	Calciudolls have the upper boundary of a calcic or petrocalcic horizon within 100 cm and are calcareous in all overlying horizons or have a texture coarser than loamy fine sand.
	Paleudolls have the upper boundary of a petrocalcic horizon within 150 cm or an argillic horizon, with some 7.5YR or redder colors, within which the maximum clay content does not decrease by more than 20% within 150 cm.
	Argiudolls have an argillic horizon.
	Vermudolls have a mollic epipedon that below 18 cm contains 50% or more (by volume) wormholes, worm casts, or filled animal burrows and rests on either a lithic contact or an underlying layer with 25% or more of similar features.
	Hapludolls—other Udolls.

redoximorphic features. They have undergone extensive iron reduction and loss due to the prolonged periods of water saturation in the presence of large amounts of organic carbon as an energy source for microbes. They have also undergone extensive melanization due to organic matter accumulation under the wet conditions. They commonly have gray and olive hues in their subsoils under a black epipedon. The Aquolls are divided into seven great groups: Cry-, Dur-, Natr-, Calci-, Argi-, Epi-, and Endoaquolls (Table 15.2).

Rendolls are found in humid regions under forests or grass and shrubs, and have formed from calcareous parent materials such as limestone-rich glacial till, chalk, and shell deposits. The mollic epipedon of Rendolls either rests directly on calcareous parent material or on a carbonate-rich cambic horizon. The mollic epipedon must be less than 50-cm thick and may be rather weakly expressed due to the dilution effect of the light-colored, calcium-rich material from which it has formed. Rendolls do not have argillic or calcic horizons. The preserving effect of the calcium-rich parent material, lack of large amounts of clay or weatherable clay minerals, and their landscape positions variously contribute to their minimal development. Though not extensive in the United States, they are of significant extent in some parts of Europe and the Yucatan peninsula of Mexico. The Rendolls are divided into two great groups: Cryrendolls and Haprendolls. Subgroups are identified on the basis of a shallow lithic contact, cryic soil temperature regime, vertic character, and presence or absence of a cambic horizon. Rendolls were classed as Rendzina soils in the previous U.S. classification (hence the derivation of their name—Rendolls).

Gelolls are the most recent suborder of Mollisols to be added to *Soil Taxonomy*, being first included in 2006 (Soil Survey Staff 2006). These soils have a gelic temperature regime in which the mean annual soil temperature ≤0°C (Soil Survey Staff 2010a). Relatively little is currently known about their distribution, but it is estimated they occupy ~155,000 km^2 globally. Gelolls are found at higher latitudes and elevation where permafrost is lacking or present below a depth of 2 m. Only one soil series within the Geloll suborder—the Kanauguk series in Alaska—is currently recognized in the NRCS database (Soil Survey Division 2010).

Cryolls are deeper, freely drained Mollisols that have a cryic temperature regime. The Cryoll suborder was incorporated into *Soil Taxonomy* in 1998 when the Boroll suborder was eliminated (Soil Survey Staff 1998). Cryolls include the soils formerly classified as Cryoborolls. Because of low temperatures, these soils typically contain relatively large quantities of organic carbon in the mollic epipedon. In the U.S., Cryolls are found mainly at higher elevations in the western states (see Figure 15.2). Cryolls are also extensive on the higher-latitude plains and mountains of Asia and Eastern Europe (Soil Survey Staff 1999). The Cryolls are subdivided into six great groups: Duri-, Natri-, Pale-, Argi-, Calc-, and Haplocryolls (Table 15.2).

Xerolls are Mollisols that have a xeric soil moisture regime, and occupy approximately 923,000 km^2 of the earth's surface (USDA-NRCS database). Those occurring in the United States have native vegetation of bunchgrass (e.g., *Festuca, Agropyron,* and *Pseudoroegneria* spp.) and shrubs (e.g., *Artemisia* spp. and *Purshia*), or savannas of grass with scattered trees. Xerolls are widely scattered throughout the states west of the continental divide. Extensive areas of Xerolls occur in California, Idaho, eastern Oregon, and eastern Washington. These soils ordinarily have a thick mollic epipedon, cambic or argillic horizon, and an accumulation of carbonates in the lower solum. The depth to secondary carbonates in Xerolls is usually much deeper than in Mollisols receiving similar amounts of annual precipitation under ustic moisture regimes (see Figure 15.3). This is testament to the greater effectiveness of leaching that occurs under a xeric moisture regime. Xerolls are also extensive in parts of Turkey, northern Africa, and some of the southern republics of the former USSR (Soil Survey Staff 1999). Xerolls have been subdivided into six great groups: Duri-, Natri-, Pale-, Calci-, Argi-, and Haploxerolls (Table 15.2). Earlier classification systems identified these soils as Brown and Chestnut soils. Data for a representative Xeroll are presented in Table 15.1.

Ustolls are the freely drained Mollisols of semiarid to subhumid climates with ustic soil moisture regimes. These are the most extensive Mollisols globally, covering over 3,900,000 km^2 or 3% of the global land area (USDA-NRCS database). Erratic rainfall occurs mostly during the growing season, and summer drought is a frequent, but erratic occurrence. Wind erosion and dust storms become problems during the drought period. The Ustolls are regionally extensive in the South American Pampas and the southern Russian steppes. They are the most extensive Mollisols in the United States, found chiefly in the southern Great Plains, where they had grass vegetation prior to cultivation. In the High Plains area of New Mexico, Texas, and Oklahoma, the Ustolls formed in regionally extensive eolian sand deposits (Holliday 1990). Most have either a Bk horizon or calcic horizon. The great groups are the Dur-, Natr-, Calci-, Pale-, Argi-, Verm-, and Haplustolls (Table 15.1). Many Ustolls were classified as Chestnut, Reddish Chestnut, and Reddish Prairie soils in earlier U.S. classification systems. Data from an Ustoll are presented in Table 15.1.

Udolls are the soils of udic moisture regimes found primarily in continental climates of the temperate regions and occupy approximately 1% of global land area (USDA-NRCS database). They are of limited extent in tropical and boreal environments. They were formed on late-Pleistocene or Holocene glacial or other deposits, under tall-grass prairie. Their well-developed mollic epipedons usually are underlain by either argillic or cambic horizons. They are very extensive in the western Corn Belt of the United States and in the humid parts of the South American Pampas. They were mostly classed as Prairie soils initially and later as Brunizems in the previous U.S. classifications. The Udolls are divided into six great groups: Natr-, Calci-, Pale-, Argi-, Verm-, and Hapludolls (Table 15.2).

Perspective

Most Mollisols formed under grass vegetation and therefore have deep, dark, and relatively fertile mollic epipedons. A few Mollisols have formed under forests, under special conditions of poor natural drainage and/or calcareous or high base status parent material. The lands where they occur are variously called prairies, llanos, steppes, and pampas. Mollisols occur predominantly in middle latitudes, but they may also occur in tropical regions as well. About 40% of the middle latitude grasslands are tall-grass prairies, and approximately 60% are short-grass steppes.

The main pedogenic process in Mollisols is melanization (darkening of the soil by addition and decomposition of organic matter). Earthworm and insect activity is higher in most Mollisols than in forest soils and plays a major role in physically mixing the soils. Accumulation of secondary calcium carbonate in subsurface horizons (Bk) is a prominent process in the drier Mollisols. Those on older landscapes in the more humid regions have argillic horizons, with evidence of lessivage (clay illuviation into B horizons).

Mollisols have been subdivided into eight suborders: Albolls, Aquolls, Rendolls, Gelolls, Cryolls, Xerolls, Ustolls, and Udolls. Humans extensively utilize these soils for wheat, corn, sorghum and millet, soybeans, pasture, and rangeland.

Oxisols: Low Activity Soils

16

Oxisols are sandy loam or finer-textured mineral soils with less than 10% weatherable minerals in the 50- to 200-micron sand fraction and low apparent cation exchange capacity (CEC) clay. If, after mixing, the surface 18 cm contains less than 40% clay, the clay content either does not increase with depth or increases so gradually that no kandic horizon is present and the upper boundary of an oxic horizon is present within 150 cm (60 in.) of the mineral soil surface, or if, after mixing, the surface 18 cm contains 40% or more clay either the upper boundary of an oxic horizon is present within 150 cm of the soil surface or the upper boundary a kandic horizon containing less than 10% weatherable minerals in the 50- to 200-micron sand is present within 150 cm (60 in.) of the surface.

Oxisol profiles have low bulk density, generally between 1.0 and 1.3 g cm^{-3}. Most horizons have strong fine and very fine granular structure. Macro pores formed between the strong granular peds provide for rapid hydraulic conductivity, much greater than normally predicted from clay content. Silt content is low and plant-available water-holding capacity is often between 0.05 and 0.15 cm cm^{-3} of horizon thickness (Buol and Eswaran 2000).

Oxisols include many of the soils previously called Laterites, Lateritic, or Latosols (Baldwin et al. 1938; Thorp and Baldwin 1940; Thorp and Smith 1949; Varghese and Byju, 1993). However, not all soils previously classed as Latosols or Lateritic (Reddish Brown Lateritic, Yellowish Brown Lateritic, etc.) can be placed in the Oxisol order. Most Ferralsols (FAO 1988) are Oxisols, but some have a kandic horizon but less than 40% clay in the surface 18 cm and classify as Ultisols or Alfisols. Many soils commonly called "red tropical or subtropical" fail the requirements of less than 10% weatherable minerals and low CEC of the clay fraction as defined by the oxic horizon and classify as Inceptisols.

Although almost all Oxisols occur between the Tropic of Cancer and the Tropic of Capricorn, it is important to recognize that many other soils are also present in that part of the world. Van Wambeke (1991) estimates that only 22.5% of the soils in the tropics are Oxisols. Oxisols are rare to absent in many tropical countries.

Setting

Most Oxisols have isothermic or isohyperthermic soil temperature regimes although there are some Oxisols with hyperthermic, thermic, or isomesic soil temperature regimes (Opdecamp and Sottiaux 1983). Oxisols are known to occur in all soil

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

moisture regimes, udic (53.0%), ustic (31.6%), perudic (11.82%), aquic (3.26%), and aridic (0.003%) (Soil Survey Staff 1999). The most extensive contiguous areas of Oxisols are in Africa and South America related to mid- to late-Tertiary geomorphic surfaces composed of material that has been subjected to surface-related weathering, eroded, transported, and re-deposited many times before coming to rest in its present location. The material within which Oxisols form on such surfaces is an accumulation of soil material composed almost entirely of minerals resistant to weathering. Some of the deposits may have stone lines of quartz and other resistant minerals, oxide-cemented gravel or a mixture of both. (See Figure 3.1.) Most often identification of individual layers is not possible because of mineral and particle size similarity. Geologically reworked and transported materials that contain appreciable quantities of weatherable minerals, such as glacially derived deposits, usually are not suitable for Oxisol formation regardless of climatic conditions (Jongen 1960; Ollier 1959; Ruhe 1956a).

Oxisols also form when easily weatherable materials, such as basic and ultra basic rocks, are exposed to warm humid conditions on stable surfaces of inactive volcanic islands and other geologically old volcanic areas (Buurman and Soepraptohardjo 1980; Beinroth 1982). These occurrences are generally of limited spatial extent and shown only on large-scale soil maps. A few Oxisols in the humid tropics and subtropics form on very stable geomorphic surfaces by slow transformation of the acid igneous rock into soil material with low activity clay over long geologic time periods.

One of the most extensive areas of Oxisols is the Sur Americana surface in central Brazil where ancient rocks were exposed to several episodes of weathering and erosion especially during the warm humid climates of Cretaceous and early Cenozoic time (Orme 2007). Most of that area has an ustic moisture regime and is vegetated with grasses and dwarf woody species commonly referred to as “cerrado.” On the same geomorphic surfaces, Quartzipsamments occupy extensive areas where the parent materials are too sandy to qualify as oxic horizons that must be sandy loam or finer texture (Lepsch and Buol 1988). It is common to have Oxisols on old fluvial terraces, pediments and other high lying erosion surfaces with Inceptisols, Ultisols, Alfisols, or even Mollisols on adjacent side slopes if mafic (basic) rock is exposed to soil formation (Figure 16.1). In Sierra Leone, Oxisols are present on young alluvial floodplains where sediment nearly devoid of weatherable minerals is being deposited (Odell et al. 1974).

Several studies in the upper Amazon basin indicate that Oxisols are not present in that area as had been anticipated from genetic considerations (Sanchez and Buol 1974). Although observations are sparse in these areas, it appears that Oxisols are confined to the lower Amazon basin and formed in material transported from the Guyana and Brazilian shields (Camargo et al. 1981; IBGE 2007). In the western portion of the Amazon basin, the sediments are derived from the Andean Mountains, contain more weatherable minerals, and Ultisols predominate with some Alfisols and Mollisols present in more basic sediments.

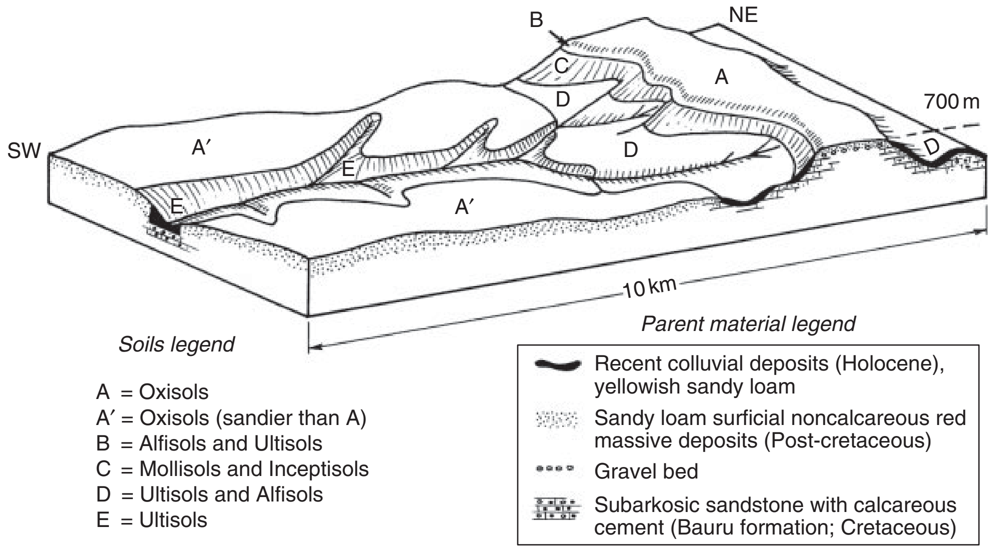

Figure 16.1. Block diagram showing distributions of soils and parent materials in the vicinity of Echapora on the Occidental Plateau, São Paulo, Brazil. (Lepsch et al. 1977a)

Oxisols support a rather wide range of vegetation including tropical rain forest, scrub and thorn forest, semideciduous forest, and grassland savanna (Mohr and van Baren 1954; UNESCO 1961; Van Wambeke 1991).

Pedogenic Processes

Desilication and concentration of aluminum and iron oxides are the major processes affecting the mineral components of well-drained Oxisols. Silicon loss is to be expected in the surface of almost all soils as rainwater infiltrates and moves through the surface layer. As infiltrated water moves downward in the soil, silicates are dissolved. The amount of silicon removed depends upon the residence time of the water around the silicate mineral, the type of silicate mineral, and the amount of surface area exposed (Wilding et al. 1977). Sand-sized quartz particles persist because of their small amount of surface area per unit weight. The loss of silicon results in almost complete decomposition of weatherable silicate minerals and 2:1 clay minerals, except the aluminum interlayer 2:1 and 2:2 intergradational (hydroxy interlayered minerals [HIM]) minerals. Kaolinite, and in some soils gibbsite and iron oxides, dominate the clay fractions of most Oxisols. Iron released from iron-bearing silicates by desilication accumulates as oxides. The citrate-dithionite extractable iron, often with considerable Al substitution, accounts for almost all of the iron in Oxisols (SMSS 1986; Fontes and Weed 1991a, 1991b).

Much desilication apparently takes place in the weathering crust near or at the rock surface in the initial material (Cady 1951). Although possibly active in present

Oxisol profiles, extensive desilication appears to have taken place during repeated cycles of weathering, erosion, and transport from soil profile to profile over vast expanses of time (Orme 2007). Once formed, the inert nature of oxic soil material precludes many other pedogenic processes. When oxic soil material is transported and deposited on younger geomorphic surfaces, it retains its properties unless mixed with weatherable minerals from other sediment sources, loess, or volcanic activity. However, significant increases in CEC and bulk density have been measured in soils formed in oxic materials on the lower portion of slopes. This has been attributed to silicon enrichment by lateral subsurface water flow during brief periods of saturation without chemical reduction (Moniz and Buol 1982; Moniz et al. 1982).

In Oxisols that have, or in the past had, a fluctuating water table relatively near the soil surface, there is often some localized iron concentration (absolute accumulation of D'Hoore 1954) to form the red-and-gray mottled redoximorphic features and/or plinthite. Plinthite-like material has been designated as laterite, first by Buchanan (1807), with later studies also using the term "lateritic iron oxide crust" (Marbut 1930; Alexander and Cady 1962; Du Preez 1949; Maignien 1959; Prescott and Pendleton 1952; Sivarajasingham et al. 1962). If large amounts of plinthite have accumulated, it tends to form a continuous phase in the soil. If the solum above the plinthite is eroded, the plinthite is subjected to repeated wetting and drying and becomes indurate nodules and concretions of petroferric material. In some cases a petroferric contact is formed. Plinthite and petroferric formation is of minor extent in Oxisols and is also present in Ultisols and other soils. Petroferric material is often a rather prominent feature where iron-rich materials are exposed on eroding escarpments. Such exposures have often led the casual observers to overestimate the extent and significance of plinthite and petroferric material in Oxisol dominated landscapes. Petroferric material may subsequently erode from the escarpments and be deposited as ironstone gravel stone lines in adjacent alluvial fans (Buol and Eswaran 2000).

Melanization and humification take place in all Oxisols. These processes are especially prominent and significant in those Oxisols with high rainfall. The high amount of biomass resulting from year-round warm temperatures also humifies and mineralizes rapidly, but a high content of organic carbon is maintained (Bennema 1974; Sanchez and Buol 1975). The organic carbon content of Oxisols is indirectly proportional to soil temperature (D'Hoore 1968). Mean organic carbon content to a depth of 1 meter in Oxisols has been found to be slightly greater than in Mollisols (Eswaran et al. 1993). Soil organic carbon contents in Oxisols have been found to have a positive linear relationship with silt plus clay contents (Lepsch et al. 1994; Zinn et al. 2007). In general, Oxisols are not as dark in color as other soils with similar organic carbon contents. It is probable that the redness of the iron oxides in Oxisol epipedons tends to mask the blackness of the organic carbon. Indeed, it is difficult if not impossible to judge the organic carbon content of Oxisols by color.

Gleization is an active process in Aquox that are saturated with water and reduced at some period or periods of time during most years. The gleization process often produces subsoil horizons with gray and red redoximorphic features that resemble plinthite but do not harden with repeated wetting and drying. Iron oxide content is low in Aquox with gray subsoil colors (SMSS 1986).

Pedoturbation appears extensive in some Oxisols. Faunal pedoturbation by insects and animals often appears to have disturbed the entire upper solum (Nye 1955; Watson 1962). The most active fauna in this process are termites that build numerous mounds using subsoil material. It should be noted that termite activity is not limited to Oxisols. The size and shape of termite mounds varies greatly throughout the world, but in the central plateau of Brazil, they characteristically are about 1 meter high and 50 cm in diameter. Termite mounds formed from poorly drained, gray-colored material clearly contrast from mounds formed from well-drained, red-colored material clearly locating the boundaries to be drawn during soil mapping.

Uses of Oxisols

Relative to most other soils, most Oxisols are chemically infertile. Their degree of infertility is reflected in the native vegetation they support. Significantly greater contents of exchangeable Ca^{2+}, Mg^{2+}, and K^{+} as well as contents of extractable P, Zn, Cu, and Mn are present in surface horizons of soils with greater density and size of woody vegetation in the cerrado of Brazil (Lopes and Cox 1977a, 1977b). Even the more highly base saturated Oxisols that support dense forests have low total quantities of calcium, magnesium, and potassium relative to most other soils. Phosphorus reacts with the aluminum and iron oxides present in the surface horizons of most Oxisols to such a degree that although total quantity may be rather great, it is very slowly available to plants.

Fast-growing crop plants seldom sustain satisfactory growth on most unfertilized Oxisols. Due to very low nutrient reserves in weatherable minerals, phosphorus retention by oxides and low CEC of Oxisols, practically all of the nutrients in the natural ecosystems are within living or dead plant tissue. On the more fertile, perhaps best stated "least infertile," Oxisols forest stands slowly accumulate plant essential nutrients in their biomass, and when the forest is cut and burned, enough nutrients are rapidly released from their organic biomass to allow subsistence farmers to harvest one or two food crops. The site is then abandoned and a natural succession of woody vegetation reestablishes because it is able to grow with a slower rate of nutrient uptake than crop plants. After several years the biomass of the natural vegetation acquires enough essential nutrients that subsistence farmers can repeat the slash-and-burn process.

On the least fertile Oxisols that naturally support only savanna grasses and sparse, stunted woody vegetation, slash-and-burn management is not possible. The scenario of events that have taken place over a 30-year period in the cerrado of Brazil provides an overview of potential Oxisol use by humans. In 1965 there was almost no human habitation in the cerrado, and farmers attempting agriculture by slash-and-burn techniques were discouraged by negligible yields (Wright and Bennema 1965). Inability to sustain indigenous human populations is a characteristic of the least fertile

Oxisols where the vegetation is so nutrient poor that large natural fauna are rare. When cattle were introduced to graze on native cerrado vegetation, their bones deteriorated from calcium and phosphorus deficiency unless supplemental mineral concentrates were supplied.

As research provided a clear understanding of the chemical limitations of Oxisols in the cerrado and in areas where economic stability enabled infrastructure necessary for commercial agriculture, an entirely different relationship of human interaction became possible. With initial applications of lime and phosphate, mixed as deeply as possible to maximize a favorable rooting depth, food crops could be successfully grown. Small amounts of zinc and copper were needed on some sites. The initial investment in rather massive amounts of phosphate fertilizer to overcome the fixation by iron and aluminum and lime to neutralize the acidity often exceeded the purchase price of the land and expense of clearing the cerrado vegetation. These initial applications had residual effects and can be viewed as capital investments. In subsequent years, annual applications of nitrogen, phosphorus, and potassium fertilizer needed to replace nutrients exported in the harvested crop and lime to maintain pH values are no greater than on any other soil growing similar crops.

By 1992, with only 10 million hectares of the estimated 204 million hectares of Oxisol-dominated cerrado cultivated, 28% of the grain production in Brazil was from the cerrado area (Lopes 1996), and beef production was rapidly increasing on fertilized pastures. With stability of markets, it became economically feasible to make initial applications of phosphate fertilizer to overcome the natural acidity and phosphate fixation capacity of the most infertile Oxisols. Facilitated by the low CEC, calcium rather rapidly moves downward within Oxisol profiles, especially when applied as gypsum, and replaces the exchangeable aluminum acidity chemically enhancing a deeper root zone (Ritchey et al. 1980).

During 8 years of soil analysis-based fertilizer and lime management on forested sites of Xanthic Hapludox soils near Manaus, Brazil, yields on 1 hectare equaled yields from 24 hectares under shifting cultivation (Cravo and Smyth 1997). Lepsch et al. (1994) monitored the long-term effects of moderate to high input farming on 77 sites, paired to represent natural and farmed Perox, Ustox, and Ustox soils in Sao Paulo, Brazil, and found no significant decreases of organic carbon in the 0–20 cm depth. Slight decreases of organic carbon, significant at the 0.05% level were found at the 60–100 cm depth in Acrudox and Acrustox soils. Significant increases in exchangeable Ca^{2+} and base saturation percentage with decreases in Al^{3+} and K^{+} were detected in the subsoil (60–100 cm) of farmed sites in all the Udox and Ustox soils.

Many Oxisols afford advantages to mechanized commercial grain production that are not available on many other kinds of soil. Many Oxisol-dominated landscapes are nearly level thus facilitating the use of large equipment. Road construction is facilitated by physical stability and rapid permeability of the strongly granular structured low activity silicate clay and iron oxide. Low indigenous populations in the cerrado region facilitated acquisition of large management units for efficient mechanized agriculture. Oxisols with ustic and udic SMRs and isothermic and isohyperthermic STRs have reliable rains during at least one growing season each year followed by a period of dryness as grain crops mature. A dry, or drier, season each year decreases the risk of grain spoilage, decreases cost of drying harvested grain, and permits mature grain to more fully dry prior to harvest allowing for maximum efficiency of harvest equipment and marketing infrastructure.

Kellogg and Orvedal (1968) clearly saw the potential of Oxisols and prophesized that they constitute the largest reserve of uncultivated soils available for development to meet world food needs. With the development of infrastructure to market harvested crops and finance lime and fertilizer purchases, farmers with both technical and business ability are now utilizing thousands of hectares of Oxisols for soybean, wheat, corn, and coffee growing and beef production on improved pastures in the cerrado of Brazil. Taking advantage of the favorable soil structure and nearly level topography, they now utilize the largest farm equipment available and most modern technologies in their successful operations.

Classification of Oxisols

Only a limited range of Oxisols are represented in the United States, and the Soil Survey Staff (1975) recognized that additional information was needed to improve the classification of Oxisols. From 1977 to 1986, an international committee for improving the classification of Oxisols (ICOMOX) collected data and involved soil classification experts around the world in workshops and field trips. In 1987 ICOMOX proposed several changes in the classification of Oxisols that were adopted in *Soil Taxonomy* (1999). Five suborders using soil moisture regime criterion are recognized in the Oxisol order: Aquox, Torrox, Ustox, Perox, and Udox (Figure 16.2).

The most extensive Oxisols have ustic or udic soil moisture regimes, the Ustox and Udox, respectively. The Perox suborder was established to group those Oxisols that in normal years experience more precipitation than potential evapotranspiration every month of the year. With a perudic soil moisture regime, it is difficult to harvest crops that require drying. The lack of a dry season hampers slash-and-burn farming systems that require a dry period for the complete burning of biomass to release adequate amounts of nutrients and obtain good crop growth. In other soil orders, soils with a perudic soil moisture regime are grouped with udic soil moisture regime soils. Soils with aquic soil moisture conditions are present within most areas of Oxisols and classified in the Aquox suborder. (See Figure 16.3.) Limited areas of Oxisols with an aridic soil moisture regime have been identified in the Torrox suborder. The wide range of soil moisture conditions serves to illustrate the significance of parent material almost devoid of weatherable minerals in the genesis of Oxisols regardless of ambient climatic conditions.

Table 16.1 outlines the great groups of Oxisols. Particular note should be taken of the *Eutr* great groups. These are Oxisols, formed in sediments derived from mafic rocks that have more than 35% base saturation (CEC_7) in all horizons to a depth of 125 cm. Some Eutrustox may have nearly 100% base saturation throughout their

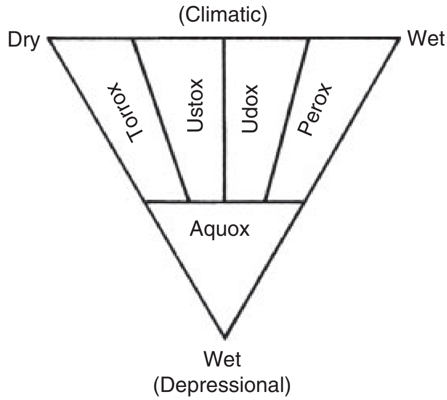

Figure 16.2. Diagram showing some relationships among suborders of Oxisols.

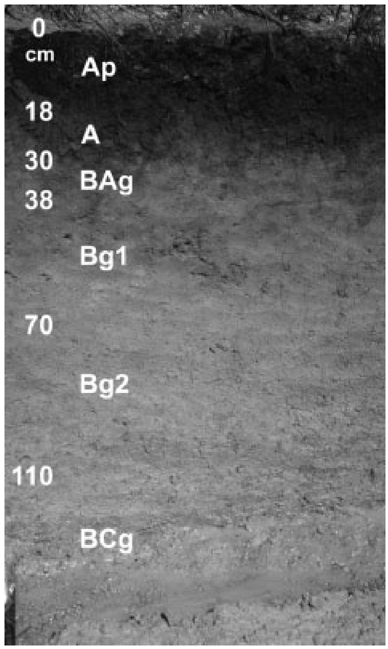

Figure 16.3. Photo of a Fine, kaolinitic isohyperthermic Aeric Haplaquox profile in the Federal District of Brazil. Characterization data from this site is published on page 658 of *Soil Taxonomy* (Soil Survey Staff, 1999). For color detail, please see color plate section.

profiles (Moura et al. 1972). Although initial crop growth may be very good on these soils, the total quantity of essential bases is limited by low CEC and near absence of weatherable minerals. Sustained harvest of crops rapidly diminishes the limited quantity of essential bases, and productivity rapidly decreases unless the soil is fertilized.

The *Acr* great groups (Figure 16.4) identify Oxisols with an apparent ECEC of less than 1.50 cmol kg^{-1} of clay in some subsoil horizon within 150 cm of the surface. Some horizons have no apparent ECEC, and some have a net positive charge. Although this may seem to be a disadvantage, research and farming practices have found low

Table 16.1. Suborders and great groups in the Oxisols order

Suborder	Great Groups
Aquox	Acraquox—apparent ECEC less than 1.5 cmol kg^{-1} clay and KCl pH of 5 or more within 150 cm of the surface
	Plinthaquox—continuous plinthite within 125 cm of the soil surface
	Eutraquox—35% or higher base saturation (CEC_7) in all horizons within 125 cm of the surface
	Haplaquox—other Aquox
Torrox	Acrotorrox—apparent ECEC less than 1.5 cmol kg^{-1} clay and KCl pH of 5 or more within 150 cm of the surface
	Eutrotorrox—35% or higher base saturation (CEC_7) in all horizons within 125 cm of the surface
	Haplotorrox—other Torrox
Ustox	Sombriustox—sombric horizon within 150 cm of the surface
	Acrustox—apparent ECEC less than 1.5 cmol kg^{-1} clay and KCl pH of 5 or more within 150 cm of the surface
	Eutrustox—35% or higher base saturation (CEC_7) in all horizons within 125 cm of the surface
	Kandiustox—more than 40% clay in surface 18 cm and the upper boundary of a kandic horizon within 150 cm of the surface
	Haplustox—other Ustox
Perox	Sombriperox—sombric horizon within 150 cm of the surface
	Acroperox—apparent ECEC less than 1.5 cmol kg^{-1} clay and KCl pH of 5 or more within 150 cm of the surface
	Eutroperox—35% or higher base saturation (CEC_7) in all horizons within 125 cm of the surface
	Kandiperox—more than 40% clay in surface 18 cm and the upper boundary of a kandic horizon within 150 cm of the surface
	Haploperox—other Perox
Udox	Sombriudox—sombric horizon within 150 cm of the surface
	Acrudox—apparent ECEC less than 1.5 cmol kg^{-1} clay and KCl pH of 5 or more within 150 cm of the surface
	Eutrudox—35% or higher base saturation (CEC_7) in all horizons within 125 cm of the surface
	Kandiudox—more than 40% clay in surface 18 cm and the upper boundary of a kandic horizon within 150 cm of the surface
	Hapludox—other Udox

ECEC in the subsoil to be an advantage because surface applied Ca^{2+} and other basic cations can rapidly move down in the soil. This encourages deeper rooting and therefore a greater supply of moisture during rainless periods that may occur during the growing season.

The *Kandi* great groups include soils with surface horizons when mixed to 18 cm that contain more than 40% clay and significant increases of clay content in the subsoil, that is, a kandic horizon that contains less than 10% weatherable minerals. The rationale for including such soils in the Oxisol order is that the initial amount of

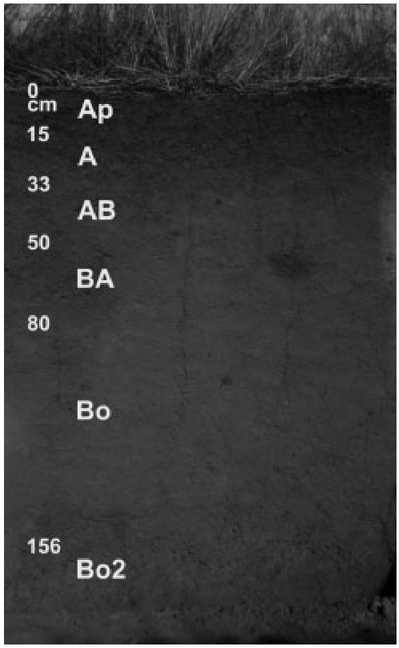

Figure 16.4. Photo of a very-fine, gibbsitic, isohyperthermic Typic Acrustox profile located in the state of Goiís, Brazil. For color detail, please see color plate section.

phosphate fertilizer needed to overcome fixation is directly related to the amount of surface area, that is, clay content in the cultivated layer. Most soils now in these great groups were formerly classified as Ultisols but are spatially associated with Oxisols and have management requirements more closely related to Oxisols than to most Ultisols.

The mineralogy in Oxisols is very restricted by the order definition, and only a limited number of mineralogy families are used in Oxisols. (See Chapter 7.)

A reaction family, Allic, is used to identify Oxisols that have more than 2 cmols of KCl extractable Al per kg soil in a 30-cm layer above 150 cm. Oxisols in Allic families have more extractable aluminum than most Oxisols and therefore require greater amounts of lime to counteract aluminum toxicity. This is an economic disadvantage when uncultivated Oxisols are first prepared for agricultural production.

Perspective

Oxisols are nonsandy mineral soils containing few weatherable minerals and low CEC. The most extensive areas of Oxisols are present in sediments that have been reworked during several cycles of erosion and deposition during which most silicate minerals, except quartz and 1:1 lattice clay, have been destroyed and oxides of iron and aluminum have been concentrated. Present soil moisture regimes range from aridic to perudic, suggesting that most Oxisols result because of the strongly weathered, nearly inert composition developed in the parent material prior to pedogenic conditions at the present site. It can be reasoned that if the parent material consists of only quartz, kaolinite, and iron oxide, few pedogenic processes are possible, and a soil formed in such material will have Oxisol properties regardless of present climatic conditions at the site.

Except for the more fertile Oxisols, that is, the Eutr great groups, most support only sparse human populations subsisting with shifting cultivation techniques and low intensity grazing. With external sources of fertilizer, highly intensive plantation agriculture (mainly sugarcane, bananas, pineapples, and coffee) has long been successful in limited areas, and there can be little doubt that with continued improvement of infrastructure, many Oxisols can become some of the most productive soils in the world. The recent expansion of soybean, wheat, corn, coffee, and beef production on Ustox and Udox in central Brazil is an example of what can be done. Clearly intensive food crop production on even the most naturally infertile Oxisols, such as the Acr great groups, is no longer limited by lack of sustainable agronomic technology. However, lack of economic infrastructure such as roads, railroads, storage facilities, and readily available technical services such as soil testing, limits production within many regions of Oxisols (Wade et al. 1988; Lopes 1996).

Spodosols: Soils with Subsoil Accumulations of Humus and Sesquioxides

17

The "white earths," the Spodosols, contrast sharply with the "black earths," the Mollisols. Spodosols encompass many of the soils called Podzols, a term from the Russian *pod* (beneath) and *zol* (ash), or "ashy underneath," referring to the light-colored E horizon (Bullock and Clayden 1980). The term "Spodosol" (wood ash) focuses attention on the spodic horizon, a subsurface horizon of illuvial accumulated organic matter complexed with aluminum, with or without iron. Spodosols are among the most eye-catching and photogenic of soils (Figure 17.1, Figure 17.2).

Setting

The major setting for Spodosols is the humid boreal climatic zone (microthermal snow-forest climate [Trewartha 1970]) where the natural vegetation was, and in many areas still is, needleleaf trees (Küchler 1970). Seasonally, rain and snowmelt water percolate through the solum. But the presence of 6 million acres (about 3 million hectares) of Spodosols in Florida (Brasfield et al. 1983) and 1.4% by area of Spodosols in the tropical lowlands of South America (Cochrane et al. 1985), counters the first impression that Spodosols are strictly boreal (and correspondingly alpine) and therefore essentially zonal (in the sense of Sibirtsev [1901]). This diversity of distribution of Spodosols indicates that coarse-textured initial material (sand, loamy sand, sandy loam) is a second common, although not universal, characteristic of the setting. A third factor is presence of vegetation that can supply mobile and sesquioxide-mobilizing organic compounds. Reports from laboratory experiments that percolates from leaves from many kinds of trees readily form miniature color "sola" in columns of sand would lead us to expect Spodosols to dominate the lands of the earth. Actually, these soils are prominent in only about 2.5% of the land area (Mokma 2006), and their extent is diminishing under the impact of human disturbance by fire, logging, cultivation, and pasturing. Hans and Jean Jenny are among persons who have established a preserve for the Blacklock soils (sandy, mixed, isomesic, ortstein, shallow Typic Duraquods) in California, lest the soil and its pigmy forest ecosystem become extinct (Jenny 1980).

Spodosols are regionally extensive near the Great Lakes of North America, in New England, eastern and western Canada, southeastern Alaska, Scandinavia, western Europe, and northwestern Russia. They occur more locally on every other continent,

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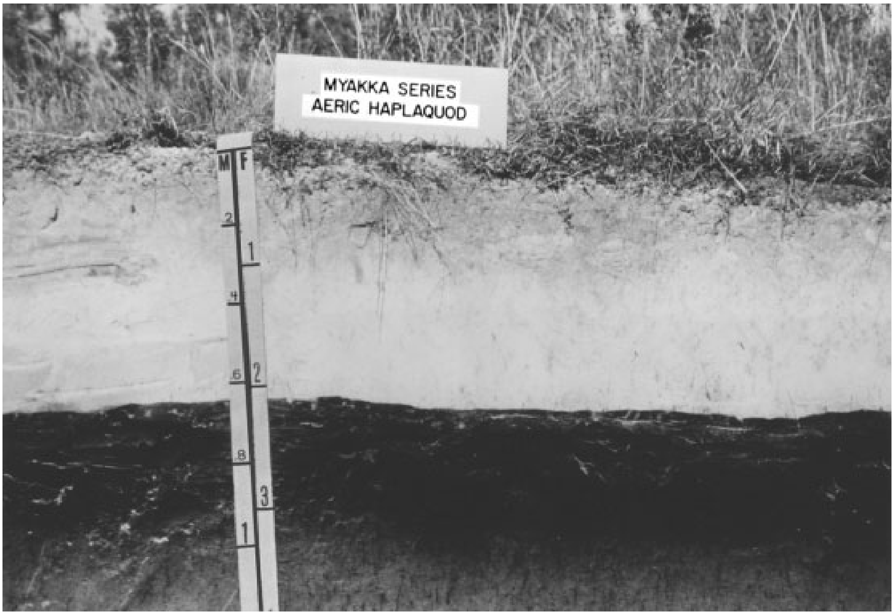

Figure 17.1. Profile of the Myakka (now classified as sandy, siliceous, hyperthermic Aeric Alaquods) soils in Florida. Scale is meters and feet.

Figure 17.2. A Haplorthod in the boreal forest of southwest Sweden formed in loess over gneissic till. The pedon is about 1 m deep. For color detail, please see color plate section.

except Antarctica, including temperate zones in the southern Andes, Tierra del Fuego, New Zealand, and Australia, and subtropical to tropical zones in Florida, Malaysia, southern Borneo, Angola, Zambia, and the tropical lowlands of South America. Favorable vegetation that supplies persistent organic compounds essential to the formation of the spodic horizon include hemlock (*Tsuga*) and other conifers (*Pinus, Agathis, Cupressus* sp.), deciduous trees (*Quercus, Fagus, Betula, Populus* sp.), heather (*Calluna vulgaris*) and other plants of the family Ericaceae, alpine grasses and sedges, tropical rain forest, savanna, and stands of palms (Bullock and Clayden 1980).

Within a given landscape, distribution of bodies of Spodosols may be fairly continuous, or it may be spotty. Where these soils effectively blanket a terrain (though with marked variability in thickness and continuity of individual horizons), a toposequence may include well to excessively drained (udic; rarely, ustic) bodies of Orthods and Humods in upland positions; and bodies of poorly drained (aquic) Aquods where pans perch the water table locally and where, in depressions, the seasonal stand of the water table is high. Scattered bodies of Spodosols reflect patchiness of (1) podzol-promoting trees (the kauri tree, *Agathis australis,* in New Zealand) (Wells and Northey 1985) or heather (*Calluna* sp. in western Europe); (2) bodies of coarse-textured geologic material in a matrix of finer-textured material; and (3) sites of local cool, moist microclimate (such as frost-prone depressions) or of a high water table.

Thickness of the solum of Spodosols varies widely. Arctic Spodosols (Gelods and Cryods) may be no thicker than 10 cm (Buurman 1984). In contrast, a Spodosol in North Carolina had a Bh horizon 9-m thick (Daniels et al. 1975). In Spodosols with a sandy particle size class, the upper boundary of the spodic horizon occurs within 200 cm of the mineral soil surface. These soils are often associated with Quartzipsamments, under which spodic horizons may lie at a depth of several meters (Andriesse 1969, 1970).

The Spodosol of North Carolina just mentioned is estimated to be more than 25,000 years old, judging by the amount of organic matter translocated down-profile annually. The Blacklock soils of California have a cemented spodic horizon (ortstein, with accessory silica cementation) and formed in deposits that are probably hundreds of thousands to a million years old (Jenny 1980), although the age of the spodic horizon itself is not certain. In glaciated terrains, Spodosols are relatively young: about 10,000 years in southwestern Sweden (Olsson and Melkerud 1989); about 2,000 to 11,300 years in Finland (Mokma et al. 2004); 3,000 to 8,000 years in Michigan (Franzmeier and Whiteside 1963a); and more than 500 years under *Tsuga canadensis* in Wisconsin (Hole 1975).

Pedogenic Processes

The dominant processes in most Spodosols are the mobilization and eluviation of aluminum and iron from O, A, and E horizons and the immobilization of these metals in short-range-order complexes with organic matter, and in some cases silica, in the

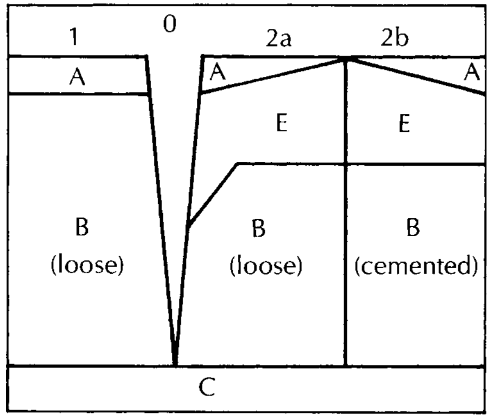

Figure 17.3. Conceptual diagram of three varieties of profiles of Spodosols differentiated by behavior of organic compounds in response to their environments (see the text).

B horizon (Mokma and Evans 2000). Collectively, these processes are known as podzolization, clearly a bundle of processes that bring about translocation of aluminum and iron under the influence of organic compounds and protons (Ponomareva 1964; Bloomfield 1954; Hallsworth et al. 1953), and, possibly, silica (Lundström et al. 2000). The process is driven largely by the production of organic acids from the decomposition of plant materials deposited on the soil surface by littering. Leaching of any carbonates present and significant replacement of the exchangeable cations Ca^{2+}, Mg^{2+}, K^{+}, and Na^{+} by H^{+} and Al^{3+} in the A horizon are prerequisite to mobilization of organic matter and sesquioxides. Malcolm and McCracken (1968) reported that a major source of mobile organic matter is tree canopy drip.

DeConinck (1980) illustrated in detail the nature and behavior of organic compounds that dominate processes of mobilization, migration, and accumulation of materials in profiles of Spodosols and other soils, particularly in cool, temperate regions. Figure 17.3 is a conceptual diagram of three kinds of Spodosols. The domain labeled O represents decomposing plant and animal remains on the forest floor, in root channels, and in animal borrows. The soil profile in column number 1 developed in the presence of supplies of Al^{3+} and Fe^{3+} abundant enough to neutralize the negatively charged organic acids by chemisorption to form bound organometallic compounds. More clay is present than is usual in Spodosols. A pronounced electrical double layer does not develop at the surfaces of organic materials, and repulsion between particles of them is not significant. The organic matter is immobilized in the O and A horizons (ochric and umbric epipedons) as a mixture of complexes and chelates, along with fragments of fresh tissues. The immobilized material does not move downward. Excess Al^{3+} and Fe^{3+} ions precipitate as hydroxides and oxyhydroxides or, if silica is present in solution, as aluminosilicates (for example, allophane or imogolite). Some of the organic matter becomes bound to the surfaces of silicate clay particles in the granular peds of the A horizon. No E horizon forms because the grains of light-colored minerals do not lose their dark coatings. Under these circumstances, spodic horizons are only weakly developed because the organic and inorganic short-range-order compounds mostly accumulate in the surface horizons, and soil properties are very similar to those of Andisols. (See Chapter 9.) The minimal spodic horizon is loose, contains iron and aluminum as free oxyhydroxides (or as Al-interlayers in clay minerals), and is transitional in form to a cambic horizon.

Profiles in columns 2a and 2b of Figure 17.3 develop in the absence of an abundant supply of Al^{3+} and Fe^{3+} ions. Parent materials are low in contents of aluminum, iron, and clay. Because the negative charge of the acidic organic compounds is not balanced by sorption of aluminum or iron, the hydrophilic organometallic compounds repel each other, disperse, and migrate. The loose, "pellet" structure in spodic horizons (profile 1 and 2a, Figure 17.3) may form by flaking of cracked illuvial humic coatings from surfaces of sand grains (Flach 1960), excretion by fauna (Robin and DeConinck 1978), or adsorption of soluble organic complexes onto surfaces of particles of clay and amorphous hydroxides (Bruckert and Selino 1978). Decay of roots and microorganisms in the B horizon is a source of organic matter in the subsoil.

Ugolini et al. (1977) found direct evidence from lysimeter studies of a Spodosol pedon of the migration of organic matter particles (0.5–1.5 μm in diameter) in the solum and of mineral particles (2–22 μm in diameter) below that. An E horizon usually forms. Below the E horizon, one of two kinds of spodic horizons may develop that have relatively high organic matter contents. In profile 2a of Figure 17.3, biologic activity keeps up with the influx of eluvial organic matter, which is converted into pellets, pedo-tubules, and porous, polymorphic (discontinuous mass with variable color and density) aggregates. Silicate mineral particles are mixed with these. Under broadleaf forest, this spodic horizon is friable to loose, high in content of relatively young humins and humic acids that can aggregate, and is well supplied with roots. It remains hydrophilic. The mean ^{14}C-residence time of organic matter is about 200 years, indicating rapid turnover. In profile 2b, under coniferous forest, the rate of eluviation of organic matter exceeds the capacity of biologic agents to rework it. Organometallic compounds accumulate as monomorphic (uniform mass) coatings (organs) that can cement the horizon (silica and aluminosilicates may be involved as well), which passes in the process from a nodular condition to ortstein (DeConinck and Righi 1983). During dehydration and shrinkage, the coatings develop minute polygonal cracks (that is, the cracked coatings, as part of the definition of spodic materials). Cementation leaves little or no room for roots, which form a mat above the cemented horizon. The horizon becomes somewhat hydrophobic. Simple, relatively old fulvic acids predominate. Mean residence time of organic matter is 10 times that in the spodic horizon of column 2a. A cemented placic horizon under anaerobic conditions in Aquods may be the result of oxygen exudation by roots in saturated soil or relict from an earlier lower stand of the water table.

Immobilization of the organometallic complexes in the spodic horizon occurs by a number of possible mechanisms (McKeague et al. 1983), but a critical factor seems to be the ratio of carbon to metal ions. When the ratio is high, the compounds are dispersed and mobile. As the ratio decreases (due to additional complexation of metal

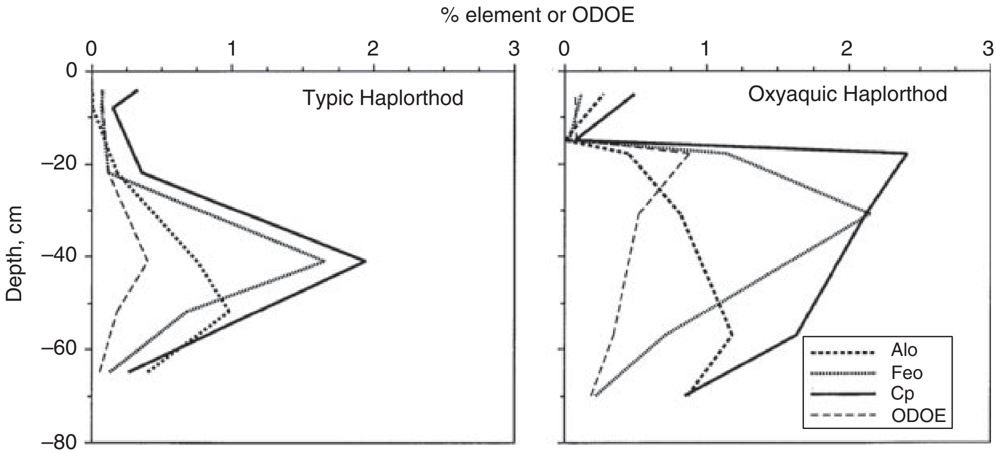

Figure 17.4. Distribution of oxalate extractable aluminum (Alo), iron (Feo), sodium pyrophosphate soluble carbon (Cp), and the optical density of the oxalate extract (ODOE) in a well-drained (Typic Haplorthod) soil and a poorly drained (Oxyaquic Haplorthod) soil formed from loess and till in southwestern Sweden. (After Southard 1994)

ions during transport; sorption by sesquioxides, existing organometallic complexes, or allophane/imogolite; or decomposition of parts of the organic component), a critical value is reached at which the complex is immobilized by precipitation or flocculation.

In addition to aluminum and iron, monovalent cations and most Ca^{2+} and Mg^{2+} ions are leached down into underlying horizons or to groundwater, along with silica (Singer and Ugolini 1974; Zabowski and Ugolini 1990), from all three kinds of profiles (Duchaufour and Souchier 1978). The translocation of silica provides a mechanism by which allophane or imogolite forms in the spodic horizon (Dahlgren and Ugolini 1989). Farmer et al. (1980) and Johnson and McBride (1989) proposed that these aluminosilicates might form in spodic horizons via transport of a hydroxyaluminum-orthosilicate complex (proto-imogolite) in solution. The capacity of some spodic horizons to sorb dissolved organic carbon may be due to the presence of imogolite (Dahlgren and Marrett 1991).

An example of the profile distributions of aluminum, iron, and carbon is shown in Figure 17.4, along with the distribution of the optical density of the oxalate extract (ODOE). Ammonium oxalate is used to dissolve short-range-order organometallic and aluminosilicate compounds (the oxalate forms strong complexes with the aluminum and iron). The carbon reported is the amount soluble in sodium pyrophosphate, which dissolves fairly labile organic matter and organometallic complexes. Generally speaking, the iron and carbon distributions parallel each other in both the well-drained and poorly drained soils, suggesting that they probably occur in complex. The aluminum content in both soils peaks at a greater depth than either the iron or carbon, suggesting the possibility that the aluminum is not in complex and may have been translocated as an inorganic ion. The ODOE parallels quite closely the carbon and iron distributions. Note that the poorly drained soil contains more total aluminum, iron, and carbon (compare the areas to the left of each depth distribution) and that the iron and carbon accumulate at shallower depths than in the well-drained soil.

Having emphasized the role of organometallic complex migration in the formation of Spodosols, it is important to recognize that other processes may be operating (Van Rompaey et al. 2007). McKeague (1981) recommended study of soils in the field, where many phenomena may be explainable in other terms (or perhaps not at all yet), as illustrated by the following examples:

1. Al^{3+} and Fe^{3+} ions are released at any depth by weathering of mineral grains.
2. A thin layer (<5 mm) placic horizon may contain inexplicably higher proportions of iron and lower proportions of aluminum than does the host spodic horizon.
3. Placic horizons also occur in Inceptisols and Histosols (DeConinck and Righi 1983; Buurman and Van Reeuwijk 1984).
4. Changing conditions and processes are indicated by two-generation cutans on sand grains in some ortsteins: an inner cutan composed of polymorphic complexes plus silt and clay, and an outer monomorphic (unlayered) organ.
5. E horizons are present in a variety of soil orders, which must mean that an array of processes is involved in their formation.
6. Much reactive Al^{3+} in the soil solution (Nilsson and Berkoist 1983) and in lower spodic subhorizons is not in organic complexes (Van Rompaey et al. 2007) and may have migrated in an inorganic complex (Childs et al. 1983; also see Figure 17.4), or perhaps a soluble hydroxyaluminum-orthosilicate complex (proto-imogolite) (Farmer 1981).
7. Microorganisms of the *Metallogenium-Siderococcus* group decompose organs but still fix Al^{3+} by forming $Al(OH)_3$ and complexes of Al with fulvic and other acids on fungal hyphae (Aristovakaya 1981). The role of iron-reducing bacteria in mobilizing iron is well known from studies of rice paddy soils and may be important in Aquods.
8. Burial of dark soil horizons, as in certain tundra areas and in tree-tip mounds in bodies of Spodosols, is an alternative process to migration of organic matter.

Thus, a large number of processes may contribute to Spodosol formation, and the relative importance of any one process may not always be clear.

A common phenomenon in Spodosol-dominated landscapes is treethrow, caused by the uprooting, generally by wind, of tall trees with shallow root systems and rigid trunks. Soil materials adhering to the tree roots are lifted and transported, producing tree-tip mounds with adjacent pits. The degree of horizon mixing is largely a function of steepness of the slope, and on steep slopes, the soil material may be completely overturned (Schaetzl 1986; Schaetzl et al. 1989). After disappearance of the tip-tree by decomposition, the mound and pit may have a relief of 1 m and length of 3 m. Growth of moss and other ground cover may prevent erosion of the mounds. Veneman et al. (1984) and Schaetzl (1986) found that the formation of E horizons was much more intensive in the pits than on the mounds, due to greater accumulation of litter and water. A major implication of treethrow is that, on the landscape scale, upper horizons of Spodosols (O, A, E, and upper B) may be very dynamic and complicate the interpretation of the rate of formation of classic Spodosol morphology.

Shifting ecotones may enhance or diminish the podzolization process (Hartshorn et al 2003). In south central Alaska forests have encroached on grasslands, converting Cryands to Cryods (Rieger and DeMent 1965). Expansion of bogs by lateral growth of sphagnum may bury Spodosols under Histosols in the taiga of Canada and southeastern Alaska. Three centuries of harvesting hay in sparsely stable-manure-fertilized alpine meadows converted Spodosols to "Brown" soils (Inceptisols) in parts of Switzerland (Bouma et al. 1969). Invasion of a *Calluna-Empetrum* heath by oak (*Quercus robur*) in Denmark caused a nearly complete conversion of Haplorthod O and E horizons into A horizons within 70 years. This conversion was accompanied by lower production of humic and fulvic acids, a less acidic A horizon, and higher rates of cellulose decomposition under oak (Nielsen et al. 1987a, 1987b). Deforestation and conversion to grass/shrub vegetation can reduce the supply of organic materials available for organometallic complex formation and result in a decrease of spodic characteristics, especially organic C contents, in spodic B horizons (Barret and Schaetzl 1998). Hole (1975) showed that the degradation of the spodic horizon is quite rapid following tree removal, with an estimated half-life of about a century.

Uses of Spodosols

Spodosols are used for forestry, pasture, hay, and cultivated crops. The soilscape in Figure 17.5 is covered for the most part with conifer-hardwood forest. Some areas in such ecosystems are devoted to recreation and others to farming (potatoes, corn silage, hay, apples). In the north central region of the United States, crop rotations on Spodosols include corn silage, oats, rye, potatoes, red clover, rutabagas, flax, strawberries, and raspberries. Commercial blueberries are grown on these soils on the coastal plain of North Carolina, in New England, and in the upper Midwest states. Fertilization and cultivation raise the nutrient levels of these soils and lead to soil compaction, to mixing of O and E horizons, and to some degradation of spodic horizons by aeration and leaching, particularly in irrigated potato fields. On the other hand, the heath lands and associated Spodosols of northern Europe persist mainly due to human activities of burning, cutting, and grazing of the heather. These activities prevent the colonization of the heath by forests (Nielsen et al. 1987a).

Spodosols and associated Quartzipsamments (in which depth to the spodic horizon exceeds 2 m) in humid tropical lowlands present problems unique to such environments. Being almost devoid of nutrients below the litter, these soils are slow to support reforestation wherever the surface horizons have been destroyed. Tropical savannas may result. The sandy E horizons of these soils are often excavated as a source of road construction material, especially in areas where gravel, rock, and plinthite materials are not available.

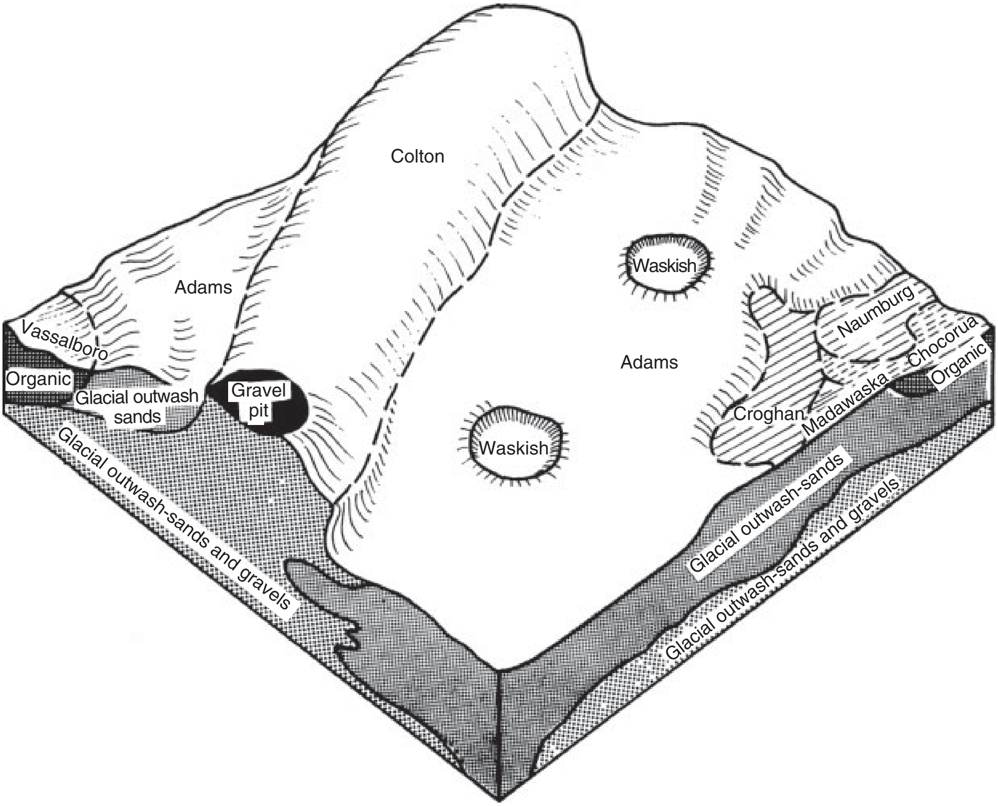

Figure 17.5. Typical pattern of soils and underlying materials in the Adams-Colton association in York County, Maine (Flewelling and Lisante 1982). The Spodosols are Adams (sandy, isotic, frigid Typic Haplorthods), Colton (sandy-skeletal, isotic, frigid Typic Haplorthods), Croghan (sandy, isotic, frigid Aquic Haplorthods), Madawaska (coarse-loamy over sandy or sandy-skeletal, isotic, frigid Aquic Haplorthods), Naumburg (sandy, isotic, frigid Typic Endoaquods). Geographically associated Histosols are Chocorua (sandy or sandy-skeletal, mixed, dysic, frigid Terric Haplohemists), Vassalboro (dysic, frigid Typic Haplofibrists), Waskish (dysic, frigid Typic Sphagnofibrists).

DeConinck (1980) recognized four classes of spodic horizons important to soil management: friable, cemented, nodular, and placic. The first is best suited to plant growth because of favorable porosity, low resistance to root penetration, and good hydraulic conductivity. The three kinds of hardened horizons may have porosities higher than those of C horizons but are resistant to penetration by roots and are barriers to upward movement of soil water. Drekker et al. (1984) found temporary ponding in slight depressions on the forest floor on Typic Haplorthods caused by crusting (plugging of soil pores by fine organic debris) of soil, and saturated conductivity (K_{sat}) was as low as 1 cm d^{-1} (per day). The cemented Bh with a K_{sat} of 8 cm d^{-1} had an actual flow rate of 32 cm d^{-1} because of a hydraulic head gradient across the spodic horizon of 3 cm cm^{-1}. Lambert and Hole (1971) reported that an ortstein spodic horizon that impeded root penetration had low hydraulic conductivity during dry periods (0.072 cm d^{-1} at −70 millibars tension), which increased under more moist conditions to 98.5 cm d^{-1} at 30 millibars tension. Deep plowing of cemented Spodosols usually improves plant growth by reducing the effects of seasonal perched water tables and resistance to root penetration, although blueberry vegetation seems to be particularly effective at promoting spodic horizon formation, and has been shown to facilitate the re-cementation of ortstein layers in Michigan Spodosols that have been deep tilled to break up the ortstein (Bronick et al. 2004).

Classification of Spodosols

The shift in emphasis from albic horizon (E) to spodic horizon (often with Bh, Bs, or Bhs horizon designations) has the advantages of (1) basing classification on a subsoil horizon less subject to alteration or removal through human activity than is the surface soil; and (2) reducing diversity of soils in the order by eliminating from it many “podzolic” soils (soils with albic horizons) of earlier classifications. On the other hand, this shift has made the separation of Andisols from Spodosols more difficult, because short-range-order aluminosilicates and organometallic complexes occur in the B horizons of soils of both orders (Nettleton et al. 1986) and causes even some rather strongly developed Podzols of other classification systems to be classified in orders other than Spodosols, particularly Entisols and Inceptisols (Mokma et al. 2004; Murashkina et al. 2005), primarily due to the spodic material color requirements.

Spodosols are identified on the basis of the occurrence of spodic materials and spodic horizons. Spodic materials have the following characteristics: a pH in water (1:1) of 5.9 or less and at least 0.6% organic carbon; and either (1) underlie an albic horizon and have either a reddish hue or are very dark gray or black (specific Munsell colors are required); or (2) with or without an albic horizon, have specific color requirements (either reddish or very dark) and are (a) cemented by organic matter and aluminum, or (b) have cracked coatings of organic matter and sesquioxides on sand grains, or (c) have at least 0.50% oxalate extractable aluminum plus one-half iron and less than half that amount in an overlying horizon, or (d) have an optical density of the oxalate extract (ODOE) of 0.25 and less than half that value in an overlying horizon. A spodic horizon is defined as an illuvial layer at least 2.5-cm thick that is not part of an Ap horizon and contains at least 85% spodic material. At the order level, Spodosols are generally defined as mineral soils that have a spodic horizon and that do not have permafrost or gelic materials.

Spodosols are subdivided into five suborders (Table 17.1, Figure 17.6), which are briefly defined as follows.

Aquods (for example, Figure 17.1) are Spodosols that have aquic conditions within 50 cm of the mineral soil surface in most years and have a histic epipedon or redoximorphic features in the upper 50 cm. These soils occupy local depressions or large areas of low relief and a high water table.

Table 17.1. Suborders and great groups in the Spodosols order

Suborders	Great Groups
Aquods	Cryaquods have a cryic soil temperature regime.
	Alaquods have less than 0.10% oxalate extractable iron in 75% or more of the spodic horizon.
	Fragiaquods have a fragipan within 100 cm.
	Placaquods have a placic horizon within 100 cm in 50% or more of each pedon.
	Duraquods have a cemented horizon within 100 cm in 90% or more of each pedon.
	Epiaquods have episaturation.
	Endoaquods—other Aquods.
Gelods	Humigelods have 6.0% or more organic carbon in a layer 10 cm or more thick in the spodic horizon.
	Haplogelods—other Gelods.
Cryods	Placocryods have a placic horizon within 100 cm in 50% or more of each pedon.
	Duricryods have a cemented horizon within 100 cm in 90% or more of each pedon.
	Humicryods have 6.0% or more organic carbon in a layer 10 cm or more thick in the spodic horizon.
	Haplocryods—other Cryods.
Humods	Placohumods have a placic horizon within 100 cm in 50% or more of each pedon.
	Durihumods have a cemented horizon within 100 cm in 90% or more of each pedon.
	Fragihumods have a fragipan within 100 cm.
	Haplohumods—other Humods.
Orthods	Placorthods have a placic horizon within 100 cm in 50% or more of each pedon.
	Durorthods have a cemented horizon within 100 cm in 90% or more of each pedon.
	Fragiorthods have a fragipan within 100 cm.
	Alorthods have less than 0.10% oxalate extractable iron in 75% or more of the spodic horizon.
	Haplorthods—other Orthods.

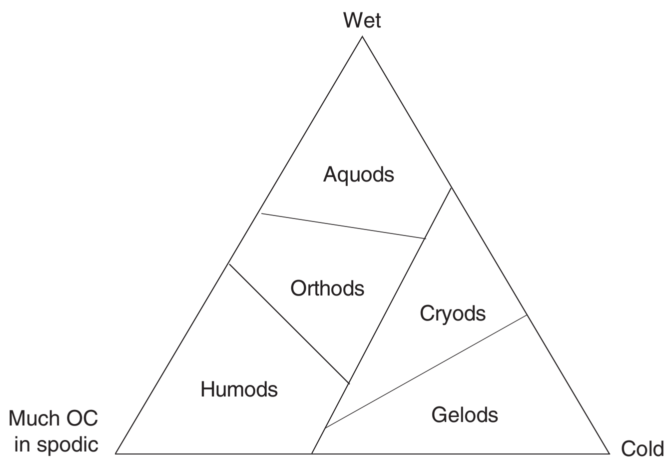

Figure 17.6. Diagram showing some relationships among suborders of Spodosols.

Gelods are better-drained Spodosols of high latitudes and altitudes that have gelic soil temperature regimes. These Spodosols occur in association with Gelisols but do not have permafrost. Soils with spodic-like horizons and permafrost are classified as Spodic subgroups of Gelisols.

Cryods have cryic soil temperature regimes (thus, are warmer than the Gelods) and generally occur at lower latitudes and altitudes than the Gelods. These are the most extensive of the Spodosols.

Humods are Spodosols that are not as poorly drained as Aquods and have spodic horizons containing 6.0% or more organic carbon.

Orthods (for example, Figure 17.2; Figure 17.3, profile 2a; Figure 17.4) are the better-drained, warmer Spodosols with spodic horizons containing less than 6.0% organic carbon.

Criteria for great groups of Spodosols are quite uniform and include fragipans, cemented horizons, placic horizons, aluminum or organic carbon content of the spodic horizon, and for the Aquods only, a cryic soil temperature regime and nature of saturation.

Perspective

The coarse texture and corresponding relatively limited total particle surface area of materials in which many Spodosols have formed make possible differentiation of contrasting E and B horizons. Many profiles observed in glaciated terrains are of young and incipient Spodosols. Some well-developed Spodosols in nonglaciated landscapes are probably steady-state systems that have maintained themselves for many thousands of years. Quantitative criteria for Spodosols select those with B horizons in which short-range-order organometallic and aluminosilicate compounds have accumulated (chiefly by illuviation) in significant amounts. Supplies of organic acids, metal ions, and silica that form the spodic horizon (generally Bs, Bh, and Bhs morphologic horizons) come from (1) decomposition of roots, surface litter, and organisms, and weathering of mineral particles, (2) mobilization of these compounds mainly in O, A, and E horizons, and (3) immobilization of the compounds in the B horizon by a number of possible mechanisms, including precipitation, chemisorption, flocculation, and decomposition or polymerization of the organic constituents. Collectively these processes are referred to as podzolization. The emphasis, in soil classification, on the spodic horizon should not divert attention from the biological, chemical, and physical dynamics of the surficial horizons (O, A, E) in natural ecosystems and the Ap horizon in cultivated areas. Low temperatures or a high water table in most Spodosols favor maintenance of relatively high organic matter contents in the spodic horizon. Cementation of portions of that horizon or an overlying albic horizon may further preserve the short-range-order materials and alter soil hydrology and the rhizosphere.

Spodosols with high water tables (or artificially drained) are placed in the suborder Aquods. Better drained Spodosols with gelic soil temperature regimes are

Gelods; those with cryic soil temperature regimes are Cryods; those organic carbon-rich spodic horizons are Humods; and those with aluminum- and iron-dominated spodic horizons are Orthods.

Spodosols are found in a wide range of ecosystems, dominantly in cold coniferous forests, but ranging into the tropical rain forests. Most are coarse, often sandy textured, and acidic to very acidic in reaction. Although often rejected for agricultural use because of their sandy, acidic condition, they are important agricultural soils in many areas.

Ultisols: Low Base Status Soils with Finer-textured Subsoil Horizons

18

All Ultisols have clay content increase with depth sufficient to identify an argillic or kandic horizon. Most well-drained Ultisols with udic, ustic, or xeric soil moisture regimes have thin A horizons, distinct light-colored E horizons, and reddish-colored argillic or kandic horizons. Poorly drained Ultisols with aquic soil moisture conditions have thicker dark-colored A horizons, often lack E horizons, and have gray-colored argillic or kandic horizons. Most Ultisols are formed under forest vegetation in parent materials containing few basic cations. Biocycling by native vegetation has concentrated basic cations in surface horizons and base saturation percentage decreases with depth.

Historically in the United States and elsewhere most Ultisols have been classified as Red–Yellow Podzolic or Reddish Brown Lateritic soils and are identified as Acrisols, Alisols, and Nitosols in the world reference base (IUSS Working Group WRB 2006). A common expression for Ultisol areas in the United States and southeastern China is “red clay hills.”

Setting

Almost all Ultisols form in acidic parent materials in locations where precipitation exceeds potential evapotranspiration during a portion of most years. Active processes of soil formation over long periods have served to deepen soil profiles while leaching and weathering the minerals present. Ultisols with udic soil moisture regimes are extensive in the southeastern United States, and Ultisols with udic, xeric, and ustic soil moisture regimes are present in the Pacific Northwest. Krebs and Tedrow (1958) have pointed out that a significant soil boundary between Ultisols to the south and Alfisols to the north exists at the terminus of glacial material in New Jersey. In piedmont physiographic areas of the southeastern United States, typical Ultisol profiles have sola 1 meter or more thick underlain by 3 or more meters of saprolite (Cr horizons) over crystalline bedrock (Pavich 1985). On alluvial fans and coastal plain sediments, sola often extend to depths of 2 or more meters (Ogg and Baker 1999; Daniels et al. 1999).

Ultisols are extensive in southeastern Asia, the upper Amazon basin in South America, the Congo River basin in Africa, and several other areas in the humid tropics where acidic parent materials are present (Sanchez and Buol 1974; Sys 1983; Lekwa

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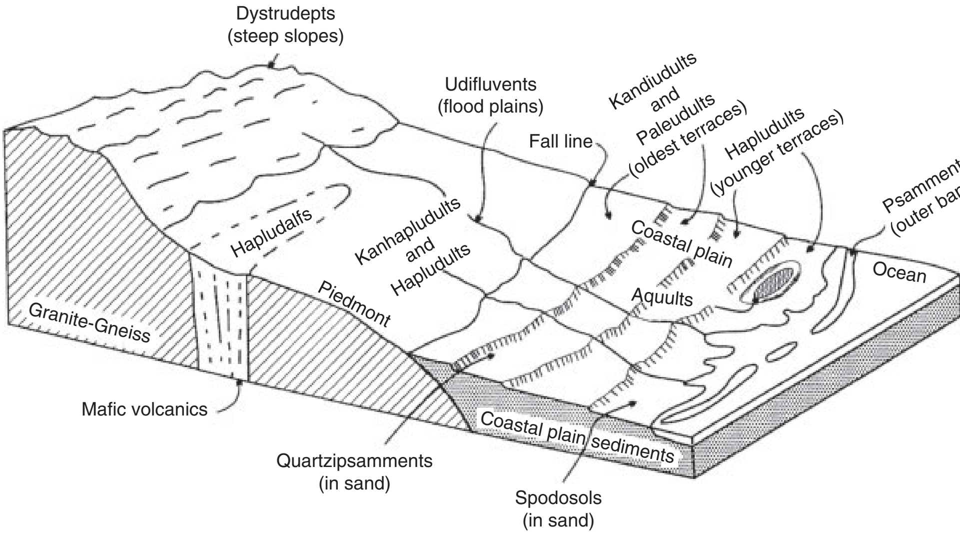

Figure 18.1. Idealized block diagram showing distribution of Ultisols and associated soils in a portion of the Carolinas, USA.

and Whiteside 1986). Most Ultisols were naturally forested, however, Ahmad and Jones (1969a, 1969b) have reported savanna vegetation on poorly drained Ultisols in northern Trinidad.

Several other kinds of soil are spatially associated with Ultisols. Where parent materials are sandy, Spodosols or Psamments are present. On steeper slopes, areas of Inceptisols, especially Dystrudepts and Dystustepts, are present, and in the recent floodplains, Fluvents and Aquents are present (Figure 18.1). Aquepts and Aquents are common associates of Aquults in poorly drained depressions.

Some Ultisols have previously been classified as Latosols or Laterites because of their red color and location in intertropical regions. In landscapes dominated by Oxisols, Ultisols are commonly formed on erosional surfaces, downslope from Oxisols (Moniz and Buol 1982; Anjos et al. 1998). Some kaolinite-dominated Bt (kandic) horizons have very low apparent cation exchange capacity, a paucity of weatherable minerals equivalent to oxic horizons, but because the A and E horizons are sandy or loamy texture, that is, they contain less than 40% clay in the upper 18 cm, they are excluded from the Oxisol order.

In acidic coastal plain sediments with low relief (Figure 3.12b), drainage catenae are present with poorly drained Aquults present in the centers of broad interfluves surrounded by Udults at the edges of the interflues where the water table is deeper (Daniels and Gamble 1967; Daniels et al. 1966c). Similar catenae are present in ustic soil moisture regimes of the upper Amazon basin with Paleustults present on edges of the interflues (Osher and Buol 1998). In rolling relief of acidic parent material, only a small percentage of the landscape is occupied by poorly drained Ultisols (Daniels et al. 1999).

Pedogenic Processes

Several individual reactions and processes are involved in the formation of Ultisols. Some seasonal leaching is present in all Ultisols. Only limited leaching is required to form Ultisols in naturally acid parent materials containing no carbonates and few weatherable minerals. Extensive leaching over a long period is characteristic of Ultisols formed in more basic parent materials. Base saturation percentage decreases with depth in Ultisols. The relative concentration of bases in A horizons suggests that biocycling by perennial tree vegetation is responsible for translocation of bases from the subsoil and depositing them as vegetative litter, this is, O horizons, on the soil surface that decomposes and is mixed into shallow A horizons. Although trees extend roots deeply into Ultisols, it is common to find the most intense proliferation of roots in the more nutrient-rich A horizons.

Extensive alteration of weatherable minerals into secondary clay minerals and oxides has taken place in many Ultisols. The clay mineral suite in Ultisols is most often dominated by kaolinite, associated with gibbsite and hydroxy-interlayered 2:1 minerals (Southern Regional Project S-14 1959). Lower apparent cation exchange capacity of the clay and thus kandic horizons are present on older, more stable geomorphic surfaces (Kleiss 1994). Muscovite mica is commonly present, probably as remnants of incompletely weathered primary minerals and tending to be more prominent in the coarse clay and silt fractions than in the finer clay fractions. Greater mica contents are most often present in Ultisols formed from mica gneiss and schist parent material (Rebertus et al. 1986). A few Ultisols formed in montmorillonite-rich sediment have montmorillonitic mineralogy (Karathanasis et al. 1986).

Lessivage, leading to the formation of argillic and kandic horizons, is very pronounced. Inability to reconstruct enough A horizon thickness to account for the large amount of clay in the argillic and kandic horizons induced Simonson (1949) to discount lessivage in Ultisols and place more emphasis on clay formation in situ in the Bt horizons. Clay formation via in situ weathering is significant in Ultisols, but the often sandy A and E horizons strongly indicate that clay eluviation has also taken place. The clay in Bt horizons clay films appears to be poorly crystalline kaolinite eluviated from the E horizon (Khalifa and Buol 1968). McCaleb (1959) postulated that clay film development in the Bt horizon was limited by the supply of weatherable minerals from which clay could form in the overlying A and E horizons. Micromorphological studies indicate that clay films in the upper Bt horizon may be destroyed and the clay released transported to form clay films in the lower Bt and upper C horizons (Brook and Van Schuylenburgh 1975; Vepraskas et al. 1996). Many Ultisols, most often those on stable upland and thus apparently the oldest profiles in a given area, do not have identifiable clay films in their argillic and kandic horizons (Gamble et al. 1970b). Lack of identifiable clay films in the kandic horizons of some Ultisol subsoils indicates that the lessivage process is relatively inactive in soils with low weatherable mineral content, although it may have been more active during earlier stages of pedogenesis (Rebertus and Buol 1985b). Pedoturbation processes appear to destroy clay films more rapidly than they form in Kandiudults and Paleudults on stable land surfaces.

Thick, sandy, arenic, and grossarenic surfaces are common in Ultisols formed from coarser-textured parent materials. In some soils, these thick sandy E horizons are the locus of spodic horizon formation. The resulting bisequal profiles with a spodic horizon overlying an argillic or kandic horizon are classified as Ultic Haplorthods and Ultic Alorthods, the overlying spodic horizon having precedence in *Soil Taxonomy*.

Ultisols are the dominant soils formed on the piedmont of the southeastern United States where a steady-state system of weathering, erosion, and isostatic uplift has operated for several million years. A typical Ultisol profile on the Piedmont consists of a 1 m or thicker solum with a sandy loam ochric epipedon and a reddish-colored, clay-textured, argillic or kandic horizon. The solum is underlain by a 1- to 10-meter-thick saprolite zone (Cr horizon) in which much of the rock structure is preserved but the density of the rock has been reduced from 2.5 g cm^{-3} to as low as 1 g cm^{-3} by isovolumetric weathering (Calvert et al. 1980a; O'Brien and Buol 1984; Buol et al. 2000). The transition from the saprolite to the argillic or kandic horizon is gradual and contains both relict rock structure and illuvial clay (Stolt et al. 1991). There are few continuous pores, and hydraulic conductive is lower in this B/C transitional horizon than in either the more clayey argillic or kandic horizon above or the underlying saprolite (Buol and Weed 1991; Schoeneberger et al. 1995; Vepraskas et al. 1996). Thinner E horizons are observed to develop in Ultisols formed from slightly basic parent rocks, such as diorite gneiss and hornblende schist saprolite, than in soils formed from granite saprolite (England and Perkins 1959). The surface soil of most well-drained Ultisols is light colored (ochric epipedon). There is usually a slight darkening of the upper 10 cm or so of most Ultisols through melanization. Solum thickness is less on slopes than on foot slope positions (McCracken et al. 1989). In tropical areas, Ultisols tend to have thinner and somewhat less distinct E horizons, containing more organic carbon and iron than do the majority of Ultisols in the southeastern United States.

Relatively high organic carbon content umbric epipdons are commonly observed in the poorly drained members of the Ultisol order, namely the Umbraquults. Under natural conditions, base saturation (CEC_7) is normally less than 50%, but many areas Umbraquults have been drained to facilitate cultivation and have for many years received applications of lime and fertilizer. The naturally occurring umbric epipedons now have a higher base saturation percentage and classify as mollic epipedons. Mollic epipedons are allowed in the Ultisols order if the underlying material has a sufficiently low base saturation status.

Two other features common to, but not definitive for, Ultisols are plinthite and fragipans. A precursor of plinthite appears to be a mottled pattern of reddish and gray colors that forms at a depth in the soil subjected to a seasonal fluctuation of the water table and often referred to as redoximorphic features (Vepraskas 1994). However, not all reddish-colored and iron-rich mottles harden irreversibly upon repeated wetting

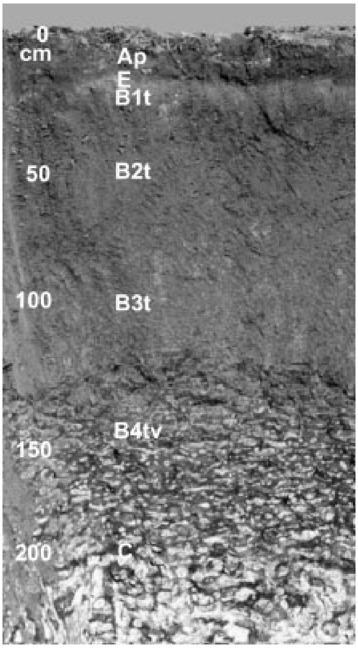

Figure 18.2. Photo of a Dothan (fine-loamy, kaolinitic, thermic Plinthic Kandiudults) profile formed in coastal plain sediments in Johnston County, North Carolina, USA.

and drying, and thus, many horizons with redoximorphic features are not plinthite (Daniels et al. 1978a). Although incipient redoximorphic features are observed in many Ultisols, only in those cases where the plinthite acts to impede drainage is it recognized in *Soil Taxonomy*. Plinthite is most often present in the subsoil of Ultisols on the most stable and hence oldest parts of the landscape (Gamble et al. 1970a, 1970b). (See Figure 18.2.)

Fragipans are found in some Ultisols, especially those with some indication of poor drainage. Fragipans, like plinthite layers, act to restrict water movement in the soil. In Ultisols, fragipans have often been confused with plinthite when gray mottles occur in a horizon of reticulate red plinthite. Both fragipans and plinthite perch water. Peds from a fragipan readily slake when dried and then submerged in water and gently agitated. Dried peds of plinthite do not slake when subjected to similar treatment (Smith and Callahan 1987). The occurrence of fragipans in Ultisols has been described by several authors (Daniels et al. 1966c; Nettleton et al. 1968; Porter et al. 1963; Soil Survey Staff 1960; Steele et al. 1969; Ogg and Baker 1999), but the genesis of fragipans remains obscure.

Uses of Ultisols

Historically, mature natural forests present on Ultisols have invited agricultural development. When native forests are cut and burned, it is usually possible to produce a few good crop yields fertilized by plant essential nutrients contained in the ashes.

As the meager supply of nutrients is removed in crop harvest, farmers either move to another location or restore nutrients with manure and/or mineral fertilizer. The low nutrient content and low base status, that is, high subsoil acidity and extractable aluminum content, of Ultisols has been, and in many areas continues to be, a major limitation to agricultural use. This limitation can be overcome by modern agricultural practices of liming and fertilization. It is necessary, however, to have adequate quantities of lime, fertilizer, and management talent available for sustained crop production. Although the immediate effect of fertilizer and lime is in the Ap horizon, increased contents of exchangeable bases and decreased acidity have been identified to depths of more than 1.5 m in Ultisols after many years of lime and fertilizer application (Buol and Stokes 1997). This results in increased rooting depth of agronomic crops (Hardy et al. 1990).

Where sustainable farming is successful via fertilization, spatial heterogeneity is common when rolling to hilly Ultisol landscapes are cultivated. On the steeper slopes, pedons have naturally thinner A and E horizons than on more level areas (McCracken et al. 1989), and cultivation often incorporates finer-textured material from the Bt horizon into the Ap horizon. The resulting finer-textured Ap horizons have lower infiltration rates, higher run-off rates, and lower crop yields than less sloping areas (Stone et al. 1985).

Timber production is extensive on many Ultisols. Compared to cultivated food crops, forests have an extremely low rate of nutrient requirement that, combined with deeper root proliferation, allows trees to grow on Ultisols too nutrient poor to sustain cultivated food crops (Buol 2008). Harvesting of native forests often leads to degradation in soil fertility, and it has been postulated that savanna vegetation has replaced forests following timber harvest in some tropical areas. Well-managed timber operations often find it profitable to fertilize Ultisols when replanting commercial timber.

Most Ultisols have relatively high contents of quartz sand and 1:1 clay that provides stable materials for earthen construction. However, in nearly level landscapes with aquic soil moisture conditions (Aquults and associated soils) in a high proportion of the area, road construction can be expensive.

Classification of Ultisols

The major diagnostic features of Ultisols are the presence of an argillic or kandic horizon and a base saturation (by sum of cations method, $CEC_{8.2}$) of less than 35% at 1.25 m (50 in.) below the upper boundary of the argillic or kandic horizon or at 1.8 m (72 in.) below the surface, whichever is deeper. If $CEC_{8.2}$ base saturation data are not available, base saturation of 50% of CEC_7 has been found to closely approximate 35% base saturation at $CEC_{8.2}$ in most kaolinite-dominated Bt horizons. If a lithic or paralithic contact is shallower than either of the above depths, the base saturation ($CEC_{8.2}$) at that contact must be less than 35%. These base saturation criteria that separate Ultisols from Alfisols are at best an arbitrary limit of doubtful genetic or agronomic significance. However, some distinction needs to be made between soils

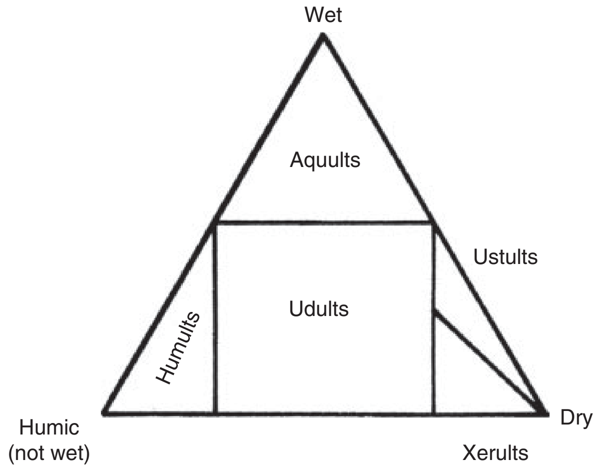

Figure 18.3. Diagram showing some relationships among suborders of Ultisols.

where the base saturation percentage decreases with depth (Ultisols) and soils where the base saturation percentage increases with depth (Alfisols), and no other criterion has been found to work as well as a specific limit of base saturation percentage at specified depths in the pedon. Base saturation criteria were placed deep enough to avoid changes in classification due to additions of lime to the Ap horizon. However, some Ultisols have been converted to Alfisols by rising sea levels and the encroachment of bases around Chesapeake Bay (Stolt and Rabenhorst 1991).

The Ultisol order is subdivided into five suborders using soil moisture regime and organic carbon content criteria (Figure 18.3).

Aquults either are saturated with water at some period of time during a normal year or are artificially drained. Because it is not practical to observe a soil throughout the year in order to classify it, certain other morphological features associated with wetness are used as criteria for defining Aquults. The criteria used are presence of redoximorphic features in all layers below 25 cm or any Ap horizon if present and above a 40-cm depth and either, (1) color chroma of 2 or less on 50% or more of ped faces, (2) chroma of 1 or less on ped faces and in the ped matrix, or (3) if in a thermic or isothermic or warmer soil temperature a hue of 2.5Y or 5Y in ped matrix within the upper 12.5 cm of the kandic or argillic horizon. Also if enough ferrous iron is present to give a positive reaction to alpha, alpha-dipyridyl within 50 cm of the surface in normal years at a time when the soil is not being irrigated identifies Aquults.

Humults have high organic carbon contents but do not have characteristics of wetness specified in the Aquults suborder. By definition, they contain more than 0.9% organic carbon in the upper 15 cm (6 in.) of the argillic or kandic horizon or 12 kg or more organic carbon in a cubic meter of the upper 1 m of the pedon, excluding O horizons. In the United States, Humults are present mainly in northwestern states where some have received surface deposits of volcanic ash and have andic-like soil properties in surface horizons. Humults may have ustic, xeric, or udic soil moisture regimes that are identified as subgroups.

Udults have a perudic or udic soil moisture regime, dry periods are of short duration, and organic carbon contents are lower than in the Humults. The water table may be in the solum for short periods each year, but aquic conditions of the Aquults are not present within 50 cm of the surface.

Ustults have an ustic soil moisture regime. Although moisture is limiting, it is seasonally available in adequate amounts for at least one crop per year.

Xerults are in areas of extremely dry summers and moist winters (xeric soil moisture regimes).

Ultisol great groups are listed in Table 18.1. Several great groups identify pedon features common to several Ultisol suborders. "Pale" great groups identify Ultisols with thick argillic horizons while "Hapl" great groups identify thinner argillic horizons. A "thick" argillic horizon is identified as having a clay content decrease of less than 20% of the maximum clay content within 150 cm of the soil surface whereas maximum clay content decreases more rapidly with depth within "thinner" argillic horizons. This criterion was selected because "thick" Bt horizons differentiates most Ultisols on the oldest geomorphic surfaces on Atlantic coastal plain sediments from "thinner" Bt horizons in most Ultisols formed on younger geomorphic coastal plain surfaces and residual rock in the piedmont physiographic region of the southeastern USA.

Ultisols with Bt horizons that have apparent CEC_7 values of 16 cmol kg^{-1} clay or less have kandic horizons. Ultisols with kandic horizon thickness equivalent to the argillic horizon thickness of "Pale" great groups were identified as "Kandi" great groups. Thinner kandic horizons are identified as "Kanhapl" (combination of "kandi" and "hapl") great groups.

"Plinth" great groups are recognized when plinthite forms a continuous layer in 50% or more of some horizon within 150 m of the surface. Plinthic subgroups are recognized when plinthite is present in more than 5% of the volume of a soil horizon within 150 cm of the soil surface. (See Figure 18.2.)

"Rhod" great groups identify Ultisols with dark red colors (hue of 2.5YR or redder and moist value of 3 or less) that have 1 unit or less value change when dry in all horizons to 100 cm. "Rhod" great groups are viewed by soil scientists as "better" than most other Ultisols. No satisfactory explanation for the distinctive color characteristics has been found, but higher than normal contents of manganese oxide are a common feature of such soils. Most have a somewhat higher base saturation percentage than other Ultisols and appear to have formed in parent materials that contain more bases.

Perspective

The most significant Ultisols features are abrupt or a clear clay content increase from the A or E horizons to kandic or argillic Bt horizons and a low base saturation percentage that decreases with depth. Ultisols are almost exclusively formed from

Table 18.1. Suborders and great groups in the Ultisols order

Suborder	Great Groups
Aquults	Plinthaquults—plinthite forms over one-half of a horizon within 150 cm of the surface
	Fragiaquults—fragipan surface within 100 cm
	Albaquults—abrupt textural change between an ochric epipedon or albic horizon and an argillic or kandic horizon that has a hydraulic conductivity of 0.4 cm hr^{-1} or less
	Kandiaquults—kandic horizon in which relative clay content decreases less than 20% within 150 cm of the surface
	Kanhaplaquults—kandic horizon in which relative clay content decreases more than 20% within 150 cm of the surface
	Paleaquults—relative clay content in argillic horizon decreases less than 20% within 150 cm of the surface
	Umbraquults—have an umbric or mollic epipedon
	Epiaquults—have episaturation
	Endoaquults—other Aquults
Humults	Sombrihumults—sombric horizon within 100 cm of the surface
	Plinthohumults—have continuous-phase plinthite in over 50% of some horizon within 150 cm of the surface
	Kandihumults—kandic horizon in which relative clay content decreases less than 20% within 150 cm of the surface
	Kanhaplohumults—kandic horizon in which relative clay content decreases more than 20% within 150 cm of the surface
	Palehumults—relative clay content in argillic horizon decreases less than 20% within 150 cm of the surface
	Haplohumults—other Humults
Udults	Plinthudults—have continuous phase plinthite in over 50% of some horizon within 150 cm of the surface
	Fragiudults—fragipan surface within 100 cm of soil surface
	Kandiudults—kandic horizon in which relative clay content decreases less than 20% within 150 cm of the surface
	Kanhapludults—kandic horizon in which relative clay content decreases more than 20% within 150 cm of the surface
	Paleudults—relative clay content in argillic horizon decreases less than 20% within 150 cm of the surface
	Rhodudults—moist color value of epipedon is 3 or less; argillic horizon has a color hue of 2.5 YR or redder with moist value of 3 or less that changes by no more than 1 unit of color value when dry
	Hapludults—other Udults
Ustults	Plinthustults—have continuous-phase plinthite in over 50% of some horizon within 150 cm of the surface
	Kandiustults—kandic horizon in which relative clay content decreases less than 20% within 150 cm of the surface
	Kanhaplustults—kandic horizon in which relative clay content decreases more than 20% within 150 cm of the surface
	Paleustults—relative clay content in argillic horizon decreases less than 20% within 150 cm of the surface
	Rhodustults—moist color value of epipedon is 3 or less; argillic horizon has a color hue of 2.5 YR or redder with moist value of 3 or less that changes by no more than 1 unit of color value when dry
	Haplustults—other Ustults
Xerults	Palexerults—relative clay content decreases less than 20% within 150 cm of the surface
	Haploxerults—other Xerults

acid (felsic) crystalline rock or acidic sediments. Ustic, xeric, udic, and perudic soil moisture regimes are present in various Ultisols depending on rainfall distribution. Aquic conditions related to the presence of shallow groundwater or perched water tables are present in the suborder Aquults.

Most Ultisols support lush forest growth, but low base saturation percentage, acidic conditions, and low contents of phosphorus limit the number of agronomic crops that can be harvested. Continuous agricultural harvests are sustainable only when lime, to overcome acidity and aluminum toxicity, and fertilizer to supply plant essential elements are routinely applied. Slash and burn (shifting cultivation) agricultural systems are common practice where lime and fertilizer are not available. Slash-and-burn agriculture, wherein one to three agronomic crops can be successfully grown immediately after nutrients accumulated during several years by slow-growing forest vegetation, are rapidly released by cutting, drying, and burning the forest vegetation. After agronomic crops are abandoned, slow-growing natural forests usually regenerate, and it takes from 15 to 30 or more years before the natural forest vegetation accumulates enough essential nutrients that it can be cut and burned to again contribute enough ash to fertilize one to three agronomic crops.

Vertisols: Shrinking and Swelling Dark Clay Soils

19

Dark, clayey soils that shrink and swell upon drying and wetting are found on every continent except Antarctica (Dudal 1963, 1965), between about 50° N and 45° S latitudes. Vertisols occupy about 320 million ha (Blokhuis 2006) or about 2.4% of the land area. Extensive areas are located in India (80 million ha), Australia (70 million ha) and Sudan (50 million ha), and the United States, China, and Ethiopia (12 to 15 million ha each). Vertisols are also locally extensive in a number of other locations including Ghana, Egypt, Chad, Cuba, Puerto Rico, Taiwan, and Uruguay (Coulombe et al. 2000; Hagenzieker 1964; Isbell 1990; Troeh 1969). In the United States, Vertisols are most extensive in Texas (6.5 million ha), South Dakota (1.5 million ha), California (1 million ha), and Montana (0.6 million ha) but are reported to occur in 25 states and territories (Coulombe et al. 2000).

Setting

A common feature in the Vertisol environments is a seasonal drying of the soil profile. Rainfall patterns associated with Vertisols are varied. Although a dry season is a necessary feature, the duration of the dry season is highly variable. The modal situation for the Vertisols involves an annual wet-dry, monsoon type (ustic) climate. The more arid Vertisol areas (Torrerts) remain dry for most of the year, with only a month or two of wetness. On the other end of the Vertisol range, soils are commonly wet (Aquerts), with moisture deficiency present for only a few weeks, often at irregular intervals, during the year.

A peculiar pedogenic landform occurs on at least 50% of the terrain occupied by Vertisols (Thorp 1957). The entire landscape may be crumpled into a complex microtopography of microknolls and microbasins (Figure 19.1). This microtopography is most commonly called “gilgai,” but is also referred to as “crabhole,” “Bay of Biscay,” “hushabye,” or “polygonal” topography. The magnitude of the microrelief appears to be greatest in the udic and ustic soil moisture regimes, and more subdued or absent in the xeric and torric regimes. Gilgai topography may take on a variety of forms at the landscape scale and has been referred to as normal, lattice, wavy, tank, stony, and melon-hole (Hallsworth et al. 1955; Hallsworth and Beckmann 1969), or lattice, dendritic, or wavy (Hagenzieker 1963).

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Figure 19.1. Gilgai landscape of microknolls and microbasins filled with water during the wet season in Texas. (From Eswaran et al. 1999)

Diapir, also called mukkara in Australia, is another term used to describe intrusions of subsoil materials through the upper layers, can be identified by a contrast in color and/or texture (Figure 19.2), and often coincides with a microknoll.

Microclimates and small-scale hydrology differ between the microrelief features of knolls and basins in a Vertisol landscape, causing differences in vegetative species, biomass production, and redox environment. The gilgai landscape also often records a complex history of past and present climates, superimposed on the local microclimates, that affected the distribution of C_3 versus C_3 plants and the dissolution, movement, and precipitation of carbonates (Kovda et al. 2003, 2006). The basins have higher humidity due to moisture release from the cracks and water ponding during wet periods, denser vegetation, and higher organic carbon contents and may be more saline than the microknolls. Mobilization of reduced Mn during the wet season may result in a significant accumulation of exchangeable Mn (Gehring et al. 1997) or of Mn-oxide nodules in the microbasins (Weitkamp et al. 1996). The knolls are drier, have higher temperatures and greater calcium carbonate contents, and are in an erosional position (Newman 1986; Wilding et al. 1990). The microrelief and distribution of soil properties are often repeated at a regular interval across the landscape. The horizontal distance from one microknoll to the next often ranges from about 3 to 10 m.

Vertisols form from a wide variety of parent material, but a common feature is a neutral to alkaline reaction. The most common parent materials include calcareous sedimentary rocks, mafic igneous rocks, volcanic ash, and alluvium from these

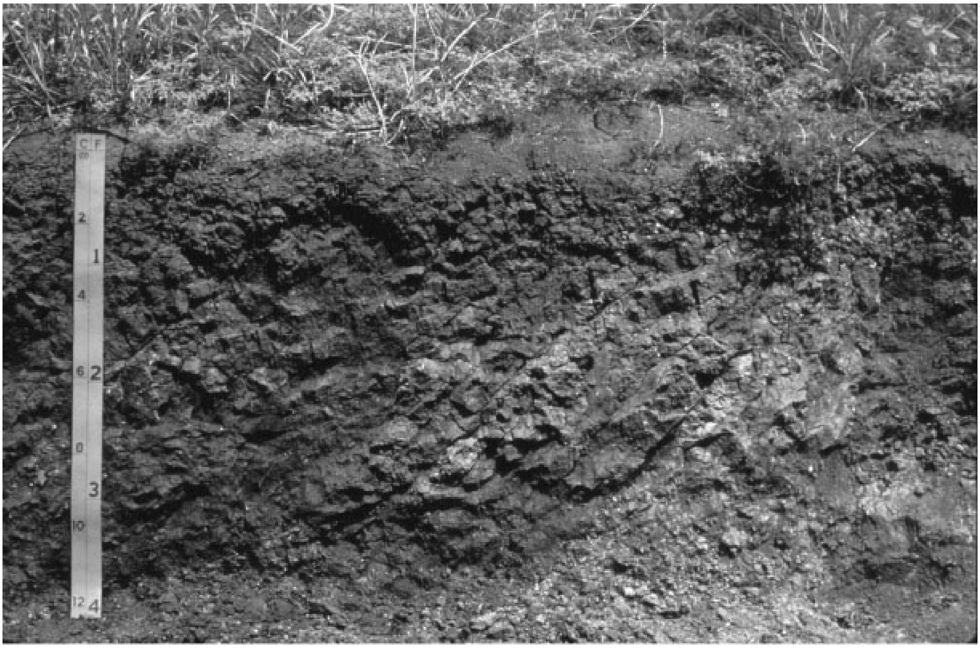


Figure 19.2. A Hapludert in Texas. The dark-colored microbasin is on the left; the lighter-colored diapir on the right creates a microknoll. The scale on the left is in decimeters and feet. (From Eswaran et al. 1999) For color detail, please see color plate section.

materials. Most Vertisols occur on nearly level to gently sloping landscapes. For example, Simonson (1954a) reported that Vertisols in India were mostly confined to landscapes with slopes from 1 to 8%. Although some Vertisols were present on steeper slopes, they were much less common on rolling landscapes, and were largely absent in hilly areas. In the Coast Ranges and foothills surrounding California's Central Valley, Vertisols formed from marine sedimentary rocks and from coarse-grained mafic igneous rocks are mapped on slopes as great as 50% (Andrews 1972; Huntington 1971).

Grasses and forbs dominate the vegetative cover on most Vertisols. Some Vertisol landscapes have shrub or woodland vegetative communities, but most large woody species are not well adapted to the shrink/swell soil properties, possibly due to root shearing and compression during drying and wetting cycles (Ahmad 1983).

Pedogenic Processes

Although there are several processes active in the formation of Vertisols, the major process seems to be shearing of wet, plastic soil materials, which may result in argillipedoturbation. To consider fully the development of the Vertisol profiles, one must first account for the high content of clay (>30% by definition) and the predominance of 2:1 expanding clay (Dixon and Nash 1968). It is not difficult to explain the presence of the necessary clay where the soils are developed from

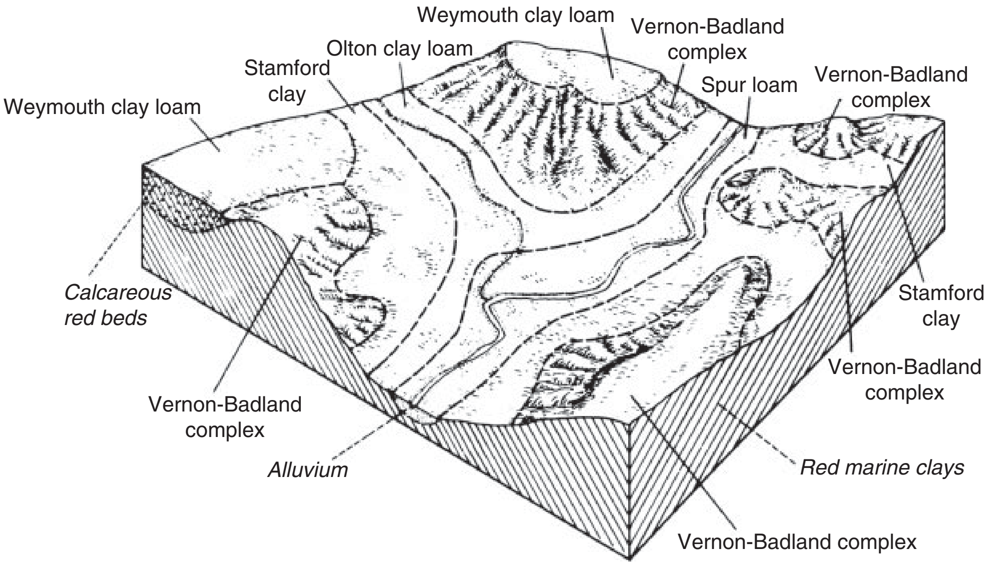


Figure 19.3. Soilscape pattern of Vertisols and associated soils in Runnels County, Texas. Soils identified are Stamford (fine, smectitic, thermic Chromic Haplusterts), Weymouth (fine-loamy, mixed, superactive, thermic Typic Haplustepts), Vernon (fine, mixed, active, thermic Typic Haplustepts), Olton (fine, mixed, superactive, thermic Aridic Paleustolls), Spur (fine-loamy, mixed, superactive thermic Fluventic Haplustolls). "Badland" identifies nonsoil land that is steep to very steep, barren land, dissected by many intermittent drainage channels. (After Wiedenfeld et al. 1970)

argillaceous limestones, marine clays (Figure 19.3), shales, or clayey, smectitic alluvium. It would appear that those Vertisols developed on basalt, however, require a fairly extensive weathering period, especially in arid to semiarid regions, unless the solum is developed from dust, volcanic ash, or colluvium deposited over the basalt.

The weathering environment of the profile must be such that the 2:1 expandable clays are not completely weathered to 1:1 clays or interlayered to the extent their expanding properties are destroyed. These environmental conditions are generally met when leaching is limited by some combination of the following: an arid or semiarid climate, a horizon or layer of slow permeability at a shallow depth below the soil (for example, a lithic contact, petrocalcic horizon, or duripan), or simply the slow permeability of the smectitic soil material itself. Under these circumstances, soil solution silica and basic cation concentrations are maintained at levels high enough for smectite to be stable, and aluminum concentrations are low, so interlayering cannot occur. Once the required content of clay and dominant 2:1 expanding clay have been achieved, shrink–swell processes begin to operate. The slow permeability and cohesiveness of the clayey, smectitic soil material ("self-preserving" properties) contribute to the preservation of paleo-Vertisols in the sedimentary rock record. The fossil Vertisols provide some of the best means for reconstructing and modeling paleo-environments (Driese 2009). Vertisols with mixed, even kaolinitic, mineralogy

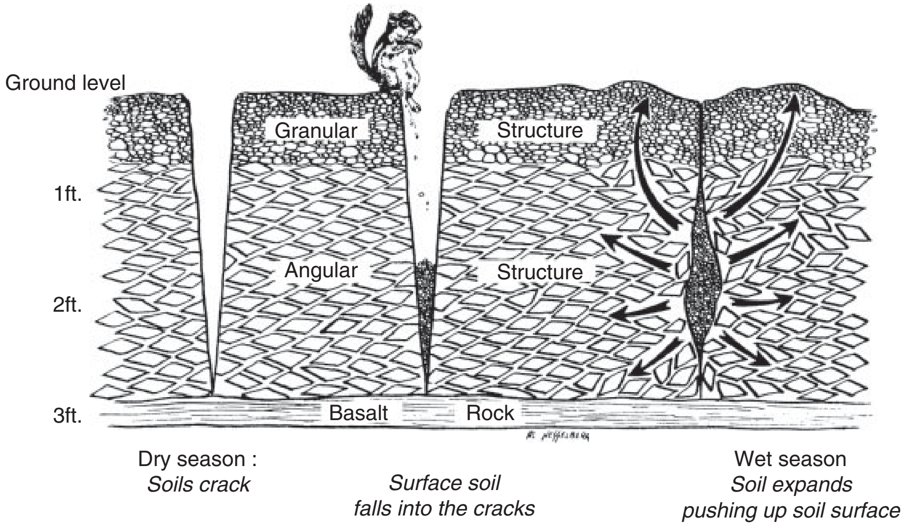


Figure 19.4. Sketch illustrating the "self-swallowing model" of Vertisols.

have been reported (Ahmad 1983; Coulombe et al. 2000), but expansible smectite, even as a subdominant component, must be largely responsible for the shrink/swell behavior.

Two major models have been proposed to account for the soil properties observed in Vertisols. The classic model, that represents processes by which the soil "inverts" itself (hence, Vertisol) is called the "self-swallowing" model, which operates in the following manner (Figure 19.4). During the dry season, the soil cracks to the surface, due to the shrinkage of the 2:1 expanding clays. The cracks often extend to a depth of 1 m or more, but cracking depth is variable and seems to be related to the depth of wetting of the profile during the wet season and the severity of drying subsequently. While the cracks are open, surface soil material falls into them. The surface material can be dislodged by several mechanisms such as animal activity, wind, or at the onset of the rainy season by water. The clays hydrate and expand on rewetting. As expansion takes place, the cracks close, but because of the "extra" material now present in the lower parts of the profile, a greater volume is required, and the expanding material presses and slides the aggregates against each other, developing a "lentil," an angular blocky structure (wedge-shaped aggregate) with slickenside features on the ped faces (Krishna and Perumal 1948). This expansion buckles the landscape, forming the gilgai microrelief. The higher organic carbon content of the microbasins may be due in part to admixtures of subsurface material into the microknoll area and slight erosion of organic-rich fines from the knolls to the basins (Templin et al. 1956). The apparently homogeneous properties, particularly soil color and clay content, of the upper parts of many Vertisols lend support to the self-swallowing model and often lead to the conclusion that the dark color is associated with organic matter derived from

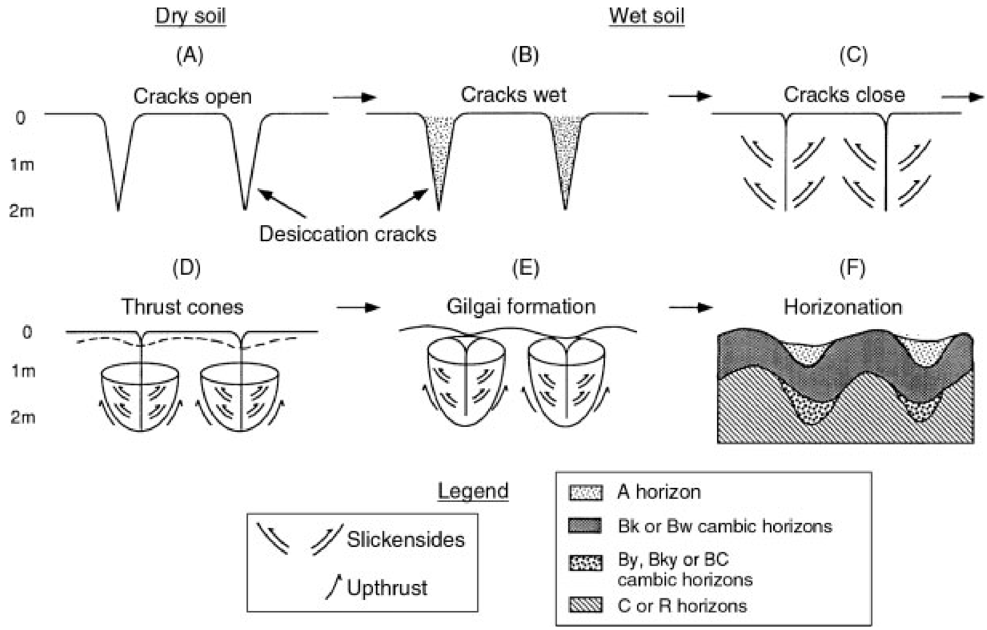


Figure 19.5. Schematic illustration of the "soil mechanics" of Vertisol development. The Bkss horizons are sometimes referred to as "vertic" horizons rather than cambic horizons. (After Coulombe et al. 1996)

incorporation of surface soil materials. Whereas the sloughing of surface materials into cracks no doubt occurs to some degree in all Vertisols, this model does not fully account for many properties of a large number of Vertisols, namely decreasing organic carbon content and increasing mean residence time of organic matter with depth and the clear differentiation of subsurface horizons with accumulations of soluble salts, gypsum, carbonates, and in some cases, even clay (Dasog et al. 1987; Wilding et al. 1990; Southard and Graham 1992).

A second model, the "soil mechanics model" (Wilding and Tessier 1988; Nordt et al. 2004), or "shear failure model" (Coulombe et al. 2000) has been proposed to explain the profile distributions of these properties (Figure 19.5). This model is based on the failure along shear planes (slickensides) of plastic soil materials when swelling pressures generated by hydration of clays exceed the shear strength of the soil material. Stress is relieved by an upward movement that is somewhat constrained by the weight of the overlying soil material, resulting in a failure shear plane that is usually inclined at 10–60° above the horizontal. This model does not require that surface material fall into cracks. Instead, subsurface material is transported upward along the slickensides to produce the microknolls of the gilgai relief, thereby exposing it to weathering and leaching processes. The slickensides, in turn, intercept percolating water and focus flow to the microbasins, where accumulation of salts, gypsum, carbonates, and Mn-oxides occurs. The presence of

Mn-oxides, produced by cycles of mild reduction, followed by oxidization and precipitation of the Mn, may be largely responsible for the dark soil colors typical of many Vertisols. Once microrelief is established, soil processes are driven largely by small-scale variations in hydrology and microclimate, and less so by pedoturbation.

It is difficult to assign all Vertisols to a similar place in a genetic scheme of soil classification. In many cases, Vertisols form from fine-textured, smectitic alluvium (Andrews 1972) and develop the characteristic cracks, slickensides, and wedge-shaped aggregate soil structure very quickly. Graham and Southard (1983) advanced another possible mode of genesis of Vertisols occurring in association with Mollisols in Utah. They concluded that some of the Vertisols there were formed when erosion removed the A horizons of Mollisols, exposing their cracking, clayey argillic horizons at the surface, leading to development of characteristics of Vertisols. Erosion of the former Mollisols is postulated as being a result of the loss of the Gambel oak trees (*Quercus gambelii*) now on the present Mollisol sites (which strongly retain surface soils against erosion) and the invasion of the present Vertisol areas by wyethia (*Wyethia amplexicaulis*). The wyethia has a long single taproot, which allows it to survive by penetrating the Vertisol clays, but the plant lacks surficial roots to hold topsoil in place. In some cases, Vertisols may be the end product of a developmental sequence involving soils whose B horizons became so clayey (for example, an argillic horizon) that shrink–swell cycles developed and eventually "swallowed" the A horizon. The high content of fine clay ($<0.2\,\mu m$) in some of these soils (Kunze and Templin 1956), and a high fine clay/coarse clay ratio, may have been the product of lessivage on a large scale.

The fate of a Vertisol may be to undergo alteration of the 2:1 clays to nonexpanding types of clay. The profile would then cease to churn as intensively, and eluviation/illuviation processes would dominate. This interpretation would suggest that Vertisols represent intermediate stages of soil development on the genetic pathway to another soil order. An alternative interpretation is that many Vertisols are stable and persistent, primarily due to dominantly low slope gradients and the slow permeability and cohesive nature of the smectitic clays. In this sense, the Vertisols may be considered to be in a steady-state, almost terminal, condition, barring a significant climatic change or major environmental disturbance. As is often the case, soils with similar properties can be produced by a number of pathways, complicating the conceptual relationships between theories of soil genesis and classification based on soil properties.

Uses of Vertisols

In general, the high content of expanding lattice clay is of primary concern in the management of these soils. Agronomic uses of Vertisols vary widely, depending on the climate and the suitability of associated soils for crop production (Probert et al. 1987). The high clay content and associated slow permeability of these soils when wet makes them desirable for cropping systems that require retention of surface water, as in the case of paddy rice cultivation. Dryland grain production is common on Vertisols in semiarid climates. Research has shown that Vertisols are fairly resilient under a variety of crop production systems but that the soils are susceptible to loss of organic carbon and to erosion during long-term cultivation and if not managed properly (Chevalier et al. 2000; Nordt and Wilding 2009). Few, if any, commercial forests or orchards are found on Vertisols due to water management difficulties and the potential for shrink-swell–induced root shearing.

Worldwide, the largest acreages of Vertisols are used for pasture. During the dry season, cracks may be wide enough to present hazardous footing for animals. Runoff from initial rains after a dry period is almost entirely infiltrated through the wide cracks. After the soil has become saturated and the cracks closed, runoff may approach 100%, making the soils susceptible to erosion, particularly if over-grazed and compacted by animal traffic (Blokhuis 2006).

Many engineering problems are associated with these soils. Structural failures are common due to differential movement of the soil surface on wetting and drying. Highways, buildings, fences, pipelines, and utility lines are moved about and distorted by the shrinking and swelling of the soil. Septic systems often fail due to slow permeability and disruption of distribution lines by shrinking and swelling. Percolation tests can be misleading and indicate too high a percolation rate if conducted during the dry season when cracks are open. It is necessary to keep these soils wet for several days to saturate and expand the clays fully to determine valid percolation values.

Classification of Vertisols

The common features of Vertisols include evidence of soil movement in the form of intersecting slickensides or wedge-shaped aggregates in a layer 25-cm or more thick within 100 cm of the mineral soil surface; clay content of at least 30% to a depth of at least 50 cm, or to a densic, paralithic, or lithic contact, duripan, or petrocalcic horizon if shallower; and cracks that open and close periodically. Although soils with slickensides and wedge-shaped aggregates often have a gilgai microtopography, this is not always the case. Further, gilgai can be partially or totally erased by cultivation, and it is sometimes possible for other microrelief features to be confused with gilgai. Therefore, gilgai surface topography is not used as a differentiating characteristic for Vertisols.

These soils have been called Grumosols (Oakes and Thorp 1950; Templin et al. 1956), Tropical Black clays or Regur (Kossovich 1912), and Tirs (del Villar 1944), as well as several other names (Simonson 1954a). It is also probable that many of these soils have been referred to as Rendzinas. The classification of Vertisols was revised significantly in 1992 to include shrink–swell soils with aquic conditions and with cryic soil temperature regimes, and to place more emphasis on diagnostic horizons and less emphasis on soil color at high categorical levels.

Because emphasis is placed on the shrink–swell characteristics of these soils at the order level, a wide array of diagnostic horizons can be included in Vertisols. Most

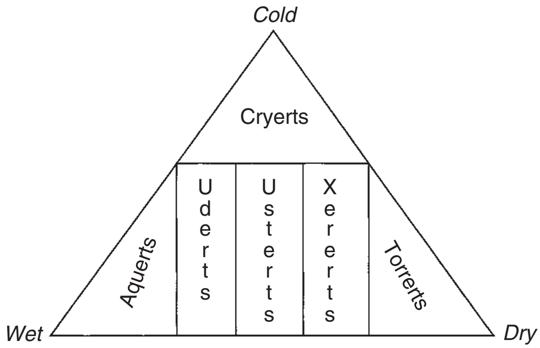


Figure 19.6. Diagram showing some relationships among suborders of Vertisols.

Vertisols have mollic epipedons (Torrerts and acidic [Dystr] great groups are common exceptions) and cambic horizons, but some have other diagnostic subsurface horizons, including argillic or natric horizons. Vertisols can have any soil temperature regime, but do not have permafrost (i.e., Gelisols).

Six suborders are recognized in the Vertisol order (Figure 19.6). They are determined on the basis of aquic conditions, cryic soil temperature regime, and on the length of time the cracks are open to the surface. Note that although the formative elements for soil moisture regimes are used in naming Xererts, Torrerts, Usterts, and Uderts, the names do not necessarily mean that the soils have those soil moisture regimes. Soil moisture regimes are poorly defined in Vertisols due to difficulties in identifying the soil moisture control section.

Aquerts are identified by aquic conditions for some time in most years and the presence of redoximorphic features. These soils may, at first, seem to violate the idea of a seasonal wet-dry cycle, until one considers that the aquic conditions need persist only long enough for some reduction and mobilization of iron and manganese to occur. Manganese oxides may be partly responsible for the dark color typical of many Vertisols. The slow permeability of clayey, smectitic soil horizons may give rise to aquic conditions in Vertisols with almost any annual precipitation pattern, although these conditions occur most commonly where soils pond water on the surface during the winter. These soils are used extensively for rice production in the Sacramento Valley of California.

Cryerts have a cryic soil temperature regime. These Vertisols are most extensive in the grassland and forest-grassland transition zones of the Canadian Prairies (Mermut et al. 1990; Mills et al. 1990), and possibly at similar latitudes in Russia.

Xererts have a thermic, mesic, or frigid soil temperature regime and, unless irrigated, have cracks that are open at least 60 consecutive days during the summer, but are closed at least 60 consecutive days during the winter. Xererts are most extensive in the western United States, primarily in California, but also occur to a limited extent in southeastern Australia and the Mediterranean region.

Torrerts, unless irrigated, have cracks that are closed for less than 60 consecutive days when the soil temperature at 50 cm is above 8°C. These soils are not extensive in the United States, and occur mostly in west Texas, New Mexico, Arizona, and

Table 19.1. Suborders and great groups in the Vertisols order

Suborders	Great Groups
Aquerts	Sulfaquerts have a sulfuric horizon or sulfidic materials within 100 cm of the surface.
	Salaquerts have a salic horizon within 100 cm of the surface.
	Duraquerts have a duripan within 100 cm of the surface.
	Natraquerts have a natric horizon within 100 cm of the surface.
	Calciaquerts have a calcic horizon within 100 cm of the surface.
	Dystraquerts have electrical conductivity less than 4.0 dS m^{-1} and pH in 0.01 M $CaCl_2$ of 4.5 or less in horizons 25 cm or more thick within 50 cm of the surface.
	Epiaquerts have episaturation.
	Endoaquerts—other Aquerts.
Cryerts	Humicryerts have 10 kg m^{-2} or more organic carbon between mineral surface and 50 cm depth.
	Haplocryerts—other Cryerts.
Xererts	Durixererts have a duripan within 100 cm of the surface.
	Calcixererts have a calcic or petrocalcic horizon within 100 cm of the surface.
	Haploxererts—other Xererts.
Torrerts	Salitorrerts have a salic horizon within 100 cm of the surface.
	Gypsitorrerts have a gypsic horizon within 100 cm of the surface.
	Calcitorrerts have a calcic or petrocalcic horizon within 100 cm of the surface.
	Haplotorrerts—other Torrerts.
Usterts	Dystrusterts have electrical conductivity less than 4.0 dS m^{-1} and pH in 0.01 M $CaCl_2$ of 4.5 or less in horizons 25 cm or more thick within 50 cm of the surface.
	Salusterts have a salic horizon within 100 cm of the surface.
	Gypsiusterts have a gypsic horizon within 100 cm of the surface.
	Calciusterts have a calcic or petrocalcic horizon within 100 cm of the surface.
	Haplusterts—other Usterts.
Uderts	Dystruderts have electrical conductivity less than 4.0 dS m^{-1} and pH in 0.01 M $CaCl_2$ of 4.5 or less in horizons 25 cm or more thick within 50 cm of the surface.
	Hapluderts—other Uderts.

South Dakota, but they are the most extensive suborder of Vertisols in Australia (Isbell 1990).

Usterts, unless irrigated, have cracks that are open for at least 90 cumulative days per year. Globally, Usterts are the most extensive suborder, encompassing the Vertisols of the tropics and monsoonal climates in Australia, India, and, Africa. In the United States, Usterts are widely distributed from Texas (Black Prairie) to Montana, and also occur in "iso-" soil temperature regimes in Puerto Rico, Hawaii, and California.

Uderts have cracks that are open less than 90 cumulative days per year and less than 60 consecutive days during the summer. These are the Vertisols of the Gulf Coastal Plain and the Black Belt in Mississippi and Alabama.

Great groups are differentiated by subsurface diagnostic horizons, sulfidic materials, reaction, organic carbon content, and nature of aquic conditions, depending on the particular suborder (Table 19.1).

A number of Inceptisols, Mollisols, Alfisols, Ultisols, Aridisols, Entisols, and Gelisols intergrade to Vertisols at the subgroup level (Vertic subgroups). These soils have significant cracking and slickensides or wedge-shaped aggregates, but not enough to be Vertisols, or have a potential linear extensibility (COLE × thickness of layer involved) of at least 6 cm in the upper 100 cm of the profile (DeMent and Bartelli 1969). At the family level, only fine (30–60% clay, for Vertisols only) and very-fine particle size classes are identified (contrasting particle size classes are also recognized). In the U.S., the vast majority of Vertisols are in the smectitic mineralogy class; a few have mixed mineralogy; three have magnesic mineralogy, and one (in Hawaii) has halloysitic mineralogy.

Perspective

Vertisols are found on every continent except Antarctica. Main areas of occurrence are in Australia, India, Sudan, Ethiopia, China, and the United States. The most extensive areas of occurrence in the United States are in Texas, South Dakota, and California. The most common environmental feature in the areas of occurrence is the seasonal drying of the soil profile. Vertisols can have any soil temperature regime but do not have permafrost. The main pedogenic process is movement and shearing of plastic soil material by the shrinking and swelling of the predominant 2:1 expanding lattice soil clays during the seasonal changes in soil moisture. In some cases, this movement causes significant mixing of soil horizons by argillipedoturbation.

Six suborders are identified based on soil temperature regime (Cryerts), aquic conditions (Aquerts), and on the duration and timing of the opening and closing of cracks (Xererts, Torrerts, Usterts, and Uderts). Great groups are identified primarily on the basis of diagnostic subsurface horizons.

Pasture is the most extensive agricultural land use. Crops grown on these soils vary widely, depending on the climate and technology for water management. Engineering problems are common due to the shrinking and swelling.

Spatial Arrangement of Soils: Soilscapes and Map Units

20

This chapter deals with the spatial distribution of the earth's soils, its components, the combination of soil and landscapes patterns we call soilscapes, the relationships between soil map units and taxonomic units, the spatial variability of soils, and how this variability may be characterized on soil maps. Much discussion focuses on the principles and concepts that have formed the basis for the soil survey program in the U.S. In addition, the development of new technologies over the past few decades has provided new ways in which to spatially represent soils and their properties, and some of these are also considered.

Definitions

The surface of earth is covered by a more-or-less regular spatial succession of different kinds of soil bodies and associated bodies of nonsoil (Hole and Campbell 1985). The soil cover pattern is "the general form of spatial distribution of soil, expressed in a definite manner by repetition of genetically linked soil areas of different taxonomic units," heterogeneous due to factors that vary over short distances (relief and parent rock). It is distinguished from the zonal or regional soil pattern that is expressed by a gradual change in soil over large areas, resulting from climatic gradients (Fridland 1976a).

A *pedon* is a specimen of a taxon in place in natural terrain. A soil individual (polypedon) is a recognizably distinct cluster of pedons that are dominated by the characteristics defined by a taxonomic unit and are separable from adjacent polypedons by boundaries that are loci of relatively abrupt changes in soil properties (Hole and Campbell 1985).

A *soil map unit* is composed of delineations on a map devised to represent spatially associated soil individuals or soilscapes, as map scales may allow. "It (a soil map unit) consists of the aggregate of all soils that are identified by a unique symbol, color, name or other representation on a map" (Van Wambeke and Forbes 1986), and much of this information may be stored as a database.

A *soilscape* is the pedologic portion of the landscape (Figure 20.1). It is an elementary soil area defined by Fridland (1976a, 1976b) as the smallest indivisible unit of the soil cover pattern.

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

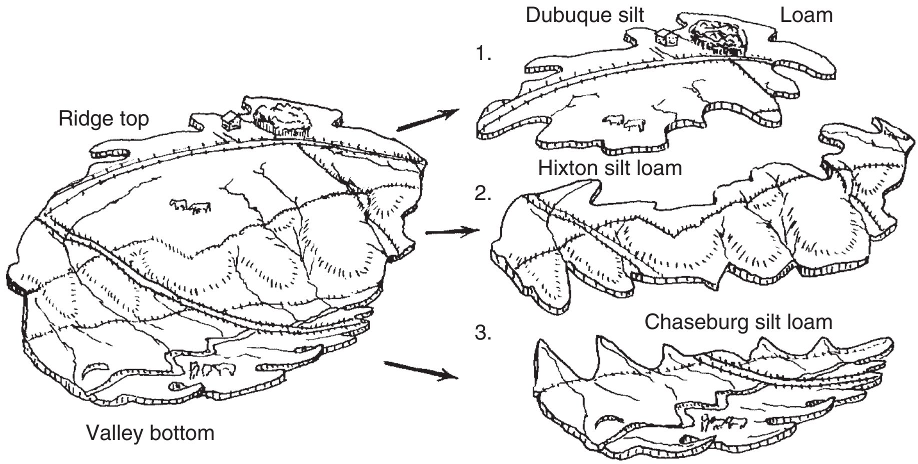


Figure 20.1. Sketch (*left*) of a sequence of soils from a ridge crest to the bottom of a small valley. This portion of a soilscape consists of three soil bodies. The Dubuque soil is a Typic Hapludalf formed in shallow leached loess over cherty residual clay over dolomite. The Hixton soil is a Typic Hapludalf formed in soft Cambrian sandstone and siltstone strata. The Chaseburg soil is a Typic Udifluvent formed in silty colluvium and local alluvium. (After Hole and Lee 1955; Soil Survey Division 2010)

Distinction between Soil Taxa and Soil Map Units

Taxonomic soil classes (taxa) can be thought of as defined portions of a multidimensional array or sets of soil properties (Arnold 1983). An individual taxonomic class (a taxon), be it a soil order or a soil series, is an abstract concept established to encompass specific ranges of soil properties as a grouping for a taxonomic categorization (Van Wambeke and Forbes 1986). Within each taxon, all included pedons are not exactly alike. This can be visualized by analogy with tree species. Trees of the same species may differ in size and shape depending on whether they are young or old, and whether they are growing in a dense forest stand or as an individual surrounded by savanna or recently buffeted by a windstorm. All taxa of soil include a similar range of properties. Each taxon is a concept with a range of properties defined by humans and not "truths" in and of themselves but our best approximations of what we perceive to be truths (Cline 1986).

A soil map unit identifies an area of land and identifies the soils present by taxonomic soil names. Every attempt to convey spatial distribution of soils via a soil map results in the inclusion of more than one pedon of the same soil class and invariably pedons of other soil classes within each delineation of that map unit. The entire range of properties of a taxon is seldom, and probably never, found within any map unit delineation. Again using the analogy to vegetation maps, a plantation of young trees is unlikely to have old mature specimens of the same species present and therefore does not include the entire range of specimens belonging to the species of tree used to identify the plantation. In addition, there may be other tree species, and perhaps some grass growing in that plantation. The fact that somewhat different pedons of the same taxon and some pedons of other taxa are included within every map unit delineation on a soil map should not be more disturbing than finding individual trees of different age, size, and morphology and perhaps a few other tree species and some grass growing within a delineation on a vegetation map.

A soil surveyor walking across the soilscape makes cartographic representations of observed soilscapes. This is done within the framework of soil taxa nomenclature (Cline 1986). Van Wambeke and Forbes (1986) state that "the fundamental difference between taxonomic units and map units is that the first is a concept resulting from subdividing the soil universe, while map units result from the grouping of soil delineations which have the same name, symbol, color or other representation." This is a convenient and simple way of differentiating taxonomic units from map units.

The search for a "basic geographic unit" of land for use as a map unit has been reviewed by Arnold (1983), indicating that such a concept functions as a link between the actual physical bodies of soil occurring in the field and the mental concepts of taxonomic classes. Arnold lists nine proposals that have been made for basic geographic units of soils. These are soil-area units (Muir 1962); soil body (Hole 1953); artificial soil body (Van Wambeke 1966); elementary soil area (Fridland 1965); land component (Gibbons and Downes 1964); component of ground surface (Van Dijk 1959); pedomorphic forms (Dan and Yaalon 1968); soil stratigraphic unit (American Commission on Strategic Nomenclature 1961); and polypedon (Johnson 1963; Soil Survey Staff 1999). To these we add the term *tessera* (Jenny 1980). Arnold (1983) concludes his listing with the statement that there is "no consensus of opinion on theoretical grounds" for defining a basic or fundamental unit of soil as a geographic body.

Over the past few decades, development of computational and information technologies—geographic information systems, global positioning systems, geostatistics, remote and proximal-data acquisition, and digital data layers—has provided many new tools for displaying the spatial arrangement of soils and their properties (McBratney et al. 2003). These new approaches, collectively referred to as *digital soil mapping*, have great potential to provide quantitative soil data that can be used for detailed environmental modeling and land management (Sanchez et al. 2009).

Although much of the soil mapping in the United States preceded this technological explosion, it is important to understand the underlying principles of soil survey as it has been conducted over the past century. This will allow the existing soil survey database to be successfully augmented with digital soil mapping methodologies to meet current and future user needs. The soil survey database for the U.S., which had traditionally been maintained in document form, was made available online via Web Soil Survey (http://websoilsurvey.nrcs.gov/) beginning in 2005. More than 95% of the nation's counties have been mapped, making Web Soil Survey the largest and one of the most detailed databases in the world.

Kinds of Soil Map Units

Maps are representations of the spatial distribution of classified objects identified at a reduced scale. Soil maps are of many scales and spatial patterns of soils. Therefore, concepts used to formulate map units differ. Map units presently used are consociations, complexes, associations, and undifferentiated groups (Soil Survey Division Staff 1993). Each is guided by concepts designed to represent the spatial distribution of pedons with maximum clarity and uniformity. The following definitions are taken from the Soil Survey Manual (Soil Survey Division Staff 1993).

Consociations are mapped areas dominated by soils of a single soil taxon and similar soils. As a rule, at least half of the pedons in a consociation are of the soil taxon for which the map unit is named. Much of the remainder of the map unit consists of soils so similar to the named soil that major use and management interpretations are not significantly different. The total area of dissimilar pedons and other components in a consociation does not exceed 25% if considered not limiting or 15% if considered limiting to major use and management interpretations. A single component of a dissimilar limiting inclusion generally does not exceed 10% if very contrasting. Percentages may vary, depending on the kind of miscellaneous area and the kind, size, and pattern of the inclusions.

Complexes and associations contain soils of two or more dissimilar taxa occurring in a regularly repeating pattern. By arbitrary rule, a *complex* is used for map units where the major components cannot be separated at a scale of approximately 1:24,000. The major components in an *association* can be separated at the same scale, but the level of detail of the survey maps is such that separation is neither practical nor needed. Two or more taxon names are used to identify complexes and associations. The amounts of other soils with limiting and nonlimiting features included are as stated in consociations.

Undifferentiated groups consist of soils of two or more taxa that are spatially associated and have the same or similar interpretations for most common land uses. Most often land use is dictated by nonsoil features such as severe flood hazard, steep slopes, or rockiness, and detailed identification of soil components serves no useful purpose. Undifferentiated map units are named for major component soils. In addition to these conceptual guidelines for naming map units, a narrative description of each map unit identifies other included soils and other features of potential significance.

Locating the Boundary of a Map Unit Delineation. The location of a soilscape boundary depends on the purpose of the map. The boundary of a delineation identifies the location where any intended user of the map may find reason(s) to alter a practice because of soil or landscape characteristics. This is illustrated in the samples of terrain shown in Figure 20.1. Farmers commonly manage their land in topographic units, and many published soil association maps group the soils in this way. The Dubuque silt loam delineation on the ridge top identifies an area of deep soils with slopes compatible with present-day practices for cultivation for row crops. The Hixton silt loam delineation identifies side slopes too steep for cultivation and best used for pasture or woodland. The Chaseburg silt loam delineation identifies a valley bottom, suitable for cultivation but subject to occasional flooding. The purpose is to identify land use potential. This leads to consociation unit delineations that are nearly homogeneous with respect to soil properties and land use possibilities, but each may contain areas of soil taxa not named by the map unit name. Most often size, extent, and spatial

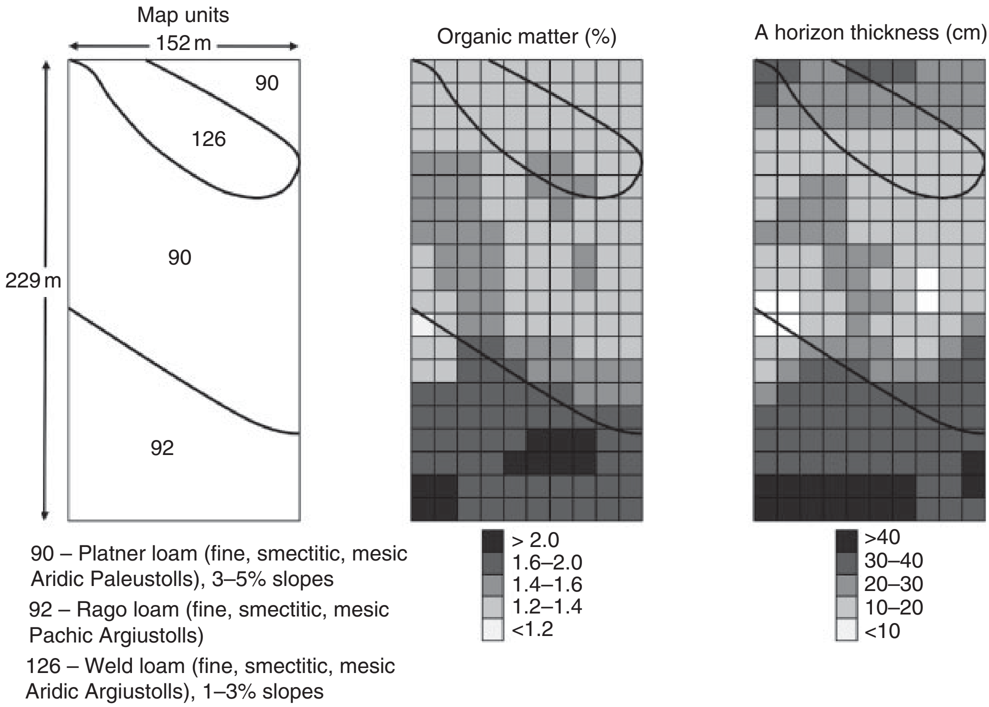


Figure 20.2 Comparison between map unit boundaries in a portion of the Logan County, CO soil survey, measured organic matter % and A horizon thickness. Grid cells are (15.24 × 15.24) m. (Adapted from Moore et al. 1993; McSweeney et al. 1994; Soil Survey Division 2010)

arrangement of included soils are not identified by the map unit name, but this information is included in the map unit description.

As illustrated in Figure 20.1, the map unit approach to represent the spatial distribution of soils on landscapes is polygon based. As such, there are limitations as to the manner and detail in which soil variability can be represented. In single-taxon map units, only the central concept of that class is represented and the polygon is portrayed as being homogeneous. In actuality, however, there likely will be variation in soil properties (Figure 20.2). In multi-taxa map units, the spatial distribution of the various and sometimes contrasting components is lost (Zhu et al. 2001). The sharp boundaries between polygons such as those shown in Figure 20.2 also portray changes in soils as a step function rather than as continuously or systematically varying across landscapes (Scull et al. 2003). Application of fuzzy logic theory to digital soil mapping can provide a means to "soften" hard boundaries and display more gradual changes in soils and their properties (Zhu et al. 2001; Hodza 2010).

Considerations of Map Scale. Every map is a reduced-scale representation of the spatial distribution of objects. A minimum area on a map is required to facilitate identification of delineations. In Table 20.1, the minimum delineation size is identified

Table 20.1. Actual land area represented by minimum size delineations (0.4 cm^2) at various map scales

Map Scale	Actual Land Area (Ha)
1:5,000,000	100,750
1:1,000,000	4030
1:500,000	1008
1:250,000	252
1:200,000	161
1:100,000	40.3
1:50,000	10.1
1:25,000	2.52
1:20,000	1.61
1:10,000	0.40

Source: Eswaran et al. 1977.

as a circle of 0.4 cm^2 on the map. This is approximately large enough to permit inclusion of letters or numbers that identify the map unit name. The actual area in hectares of land identified by this minimum delineation size is presented for various map scales in Table 20.1.

Orders and Standards of Soil Surveys

Some users of soil surveys need very specific and detailed information about the nature of soil resources. For these potential users, the information needed is about the nature of soil areas of a few hectares (or acres) or less. Other users may need only a broader depiction of soil areas of interest to them, such as areas of thousands of hectares (acres) each. Different levels of detail are therefore provided by the soil survey maps. Map scale and levels of detail are arranged in classes of soil surveys called orders of soil surveys (Soil Survey Division Staff 1993, pp. 47–56). Orders of soil survey maps differ in kinds of map units and levels of soil taxa used for identifying the map units, for example, named as soil series, families, subgroups, great groups, suborders, orders, and phases of these taxonomic entities.

Standards of purity for the mapped soil areas are attained by adjusting the mapping operations according to the precision required. If the standards require that areas of limiting dissimilar soils as small as 0.1 ha (0.25 acre) be delineated, the area must be mapped at a scale large enough to permit identification of such small areas and the soil must be examined in enough places to find them. The five orders of USDA soil surveys are classified in Table 20.2. The level of detail of these survey orders can also be generally related to grid-cell sizes of digital soil maps; these range from (5×5) m to $>(2 \times 2)$ km (McBratney et al. 2003).

First and Second Order Soil Mapping. Almost all first and second order soil mapping is done in the field. After obtaining a base map of appropriate scale, most

Table 20.2. Kinds of soil surveys

Mapping Order and Intended Use	Field Procedures	Minimum Size Delineation (hectares)	Typical Components	Kinds of Map Units	Appropriate Scales
First-order: Experimental plots and building sites	Delineation boundaries identified by frequent boring	1 or less	Phases of series and miscellaneous areas	Consociations, complexes, and miscellaneous areas	1:15,840 or larger
Second-order: General agricultural and urban planning	Remote sensing withfield observation of each delineation	0.6–4	Phases of series and miscellaneous areas	Consociations, complexes; few associations	1:12,000 to 1:31,680
Third-order: Range management; community planning	Delineations plotted from remote sensing; then transecting	1.6–16	Phases of series or higher taxa	Associations, complexes, consociations; undifferentiated groups	1:20,000 to 163,360
Fourth-order: Broad land use potential identification	Remote sensing; some transects	16–252	Phases of series or higher taxa	Associations, complexes, consociations; undifferentiated groups	1:63,360 to 1:250,000
Fifth-order: Broad geographical teaching	Assimilated data; some transecting	252–4000	Phases of higher taxa	Associations	1:250,000 to 1:1,000,000 or smaller

Source: Soil Survey Division Staff 1993.

often aerial photographs that provide for stereoscopic coverage or detailed topographic maps, the soil surveyor sketches an outline of landforms and tonal patterns that are visible. Stereoscopic observations of aerial photos are especially useful in many, but not all, areas. With a copy of the base map suitable for transport and weather proofed to avoid inclement conditions, the surveyor traverses the land observing vegetative cover, land forms, and progressively examining the soils by boring auger holes as determined by "free survey" techniques. Delineations of landforms and tonal patterns previously sketched are altered by direct observation as the surveyor traverses the terrain. In the process of mapping, the field soil scientists utilize knowledge of soil genesis principles, combining observations of surface soil and subsoil materials at sites of auger holes with observations of topography and vegetative cover. No specific number of borings is required by "free survey" technique, and the location of the borings is determined by judgment. Experienced soil scientists accurately delineate map units by directly examining less than one-thousandth of the soil below the surface (Hudson 1992). When first mapping an area, the soil scientist will frequently examine soil profiles to establish the association of pedon properties with the landscape and vegetative features observed. After landscape and vegetative associations with pedon properties have been firmly established, the frequency of profile examination decreases and is concentrated in areas where such associations are less predictable.

Soil scientists are aware that all soilscapes are assemblages of many pedons on the land. During the process of mapping, they depict these assemblages as best they can, within the limits of the scale of the map. More "ground truth" information is encountered than can be recorded on the map, which regardless of scale is a generalized picture of nature. Some of the spatially minute detail may be rescued from omission by the use of special symbols to indicate sandy, stony, wet, or hummocky conditions at particular sites and by taking notes that may be summarized as narrative in the map unit descriptions. Farmers and other indigenous individuals are queried regarding crop yields and other management related responses concerning the behavior of different soils in extreme weather conditions. Although such data must be considered anecdotal, it alerts the surveyor to conditions not seen during the brief time of survey and when aggregated provides useful information. Also, the inclusion of locally accepted descriptions of various aspects of soils and their management in the text of the soil survey report increases indigenous understanding and support of the soil survey.

Generalized and Schematic Mapping for Third, Fourth, and Fifth Order Soil Maps. Small-scale soil maps are used not only for planning, zoning, and legislative decisions but also for ecological research and environmental protection studies. Schematic maps are prepared by collecting and organizing information on patterns of soil-forming factors obtained from remotely sensed information. Broad delineations are outlined on appropriate base maps, and the compiler makes a preliminary estimate of the probable soil pattern. A small number of delineations of similar units are then selected for random or systematic soil observation. Systematic maps are most useful only where more detailed soil surveys are not available.

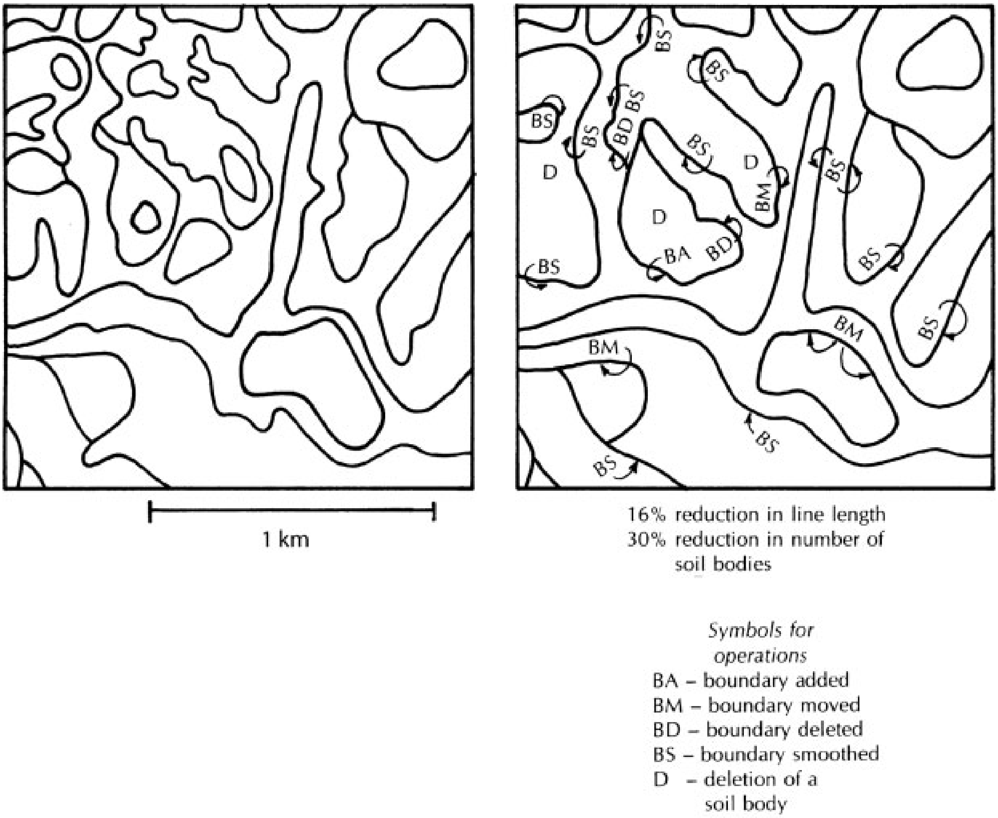


Figure 20.3. Example of graphic generalization (*right*) of a detailed soil map (*left*). (Map based on part of map sheet No. 9, Bartelme 1977)

Generalized soil maps are not derived directly from field mapping but are simplifications of a known map unit distribution (Hole and Campbell 1985). They are most often created by reducing complex cartographic detail of order 1 and 2 soil surveys to smaller-scale maps that depict broad regional patterns of soil distribution. Generalized soil maps are prepared by reorganization and omission of some information from more detailed primary source maps. The generalization process for map compilation can take one of four forms or strategies: graphic, taxonomic, spatial, and sampling generalization (Hole and Campbell 1985).

Graphic Generalization. This consists of adjusting shapes of line segments that separate delineations, yet leaving the original map legend intact or nearly so. Graphic generalization can be carried out in one or more operations on the source map (Figure 20.3): boundary smoothing, deleting an entire delineation by removing its boundaries, enlarging small delineations, shifting boundary locations, and adding boundaries. No technical knowledge of soils and their distribution is required because the operation is mechanical—done on the manuscript map or with a computer and monitor. Practically all generalized soil maps have had some graphic generalization.

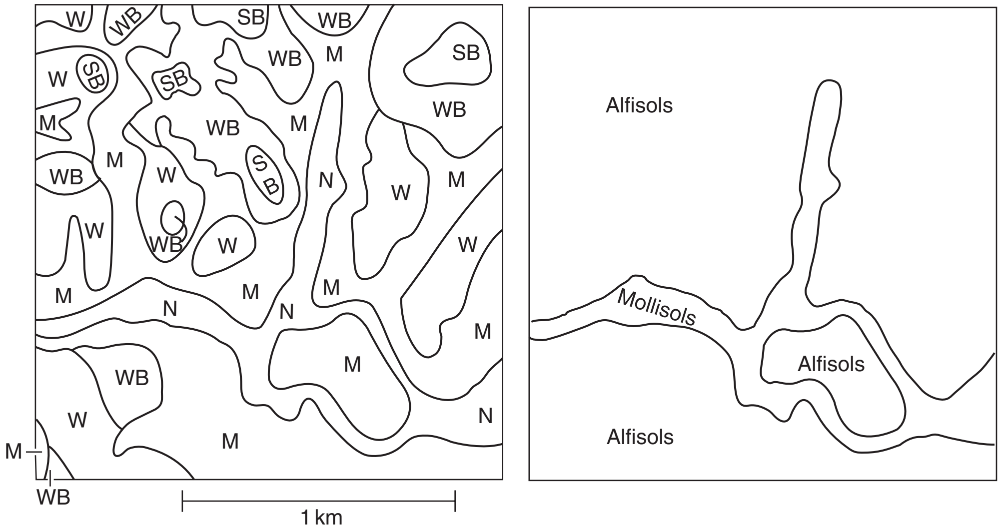


Figure 20.4. Taxonomic generalization (*right*) of a detailed soil map (*left*). (Map based on part of map sheet no. 9, Bartelme 1977)

SB = Santiago silt loam. Coarse-loamy, mixed, superactive, frigid Haplic Glossudalfs.
W, WB = Withee silt loam. Fine-loamy, mixed, superactive, frigid Aquic Glossudalfs.
M = Marshfield silt loam. Fine-loamy, mixed, superactive, frigid Mollic Epiaqualfs.
N = Mann silt loam. Fine-loamy, mixed, superactive, frigid Typic Epiaquolls.

This strategy is often used in combination with one or more of the other strategies of generalization to create delineations suitable for a smaller-scale map. Graphic generalization is deleterious to precision of boundary locations if done at the same scale as the original map was produced.

Taxonomic Generalization. This includes deletion of boundaries between adjacent delineations that are taxonomically similar. This can be accomplished in two ways: either by combining map units that are taxonomically similar in a hierarchal classification system such as Soil Taxonomy (Figure 20.4) or by use of statistical multivariate measurements to find those units showing the most similarity and combining them. This type of generalization produces more broadly defined map units while not detracting from accuracy but lowering the precision of soil property information. This type of generalization is similar in many respects to the categorical generalization described by Orvedal and Edwards (1941).

Spatial Generalization. This involves combining map units that are consistently found adjacent to one another, whether taxonomically similar or not. This is generally done for land management or technically similar uses. It is accomplished by deletion of boundaries between adjacent map units, not randomly or mechanically, but with some technical knowledge of soilscape patterns of the region and appreciation of the

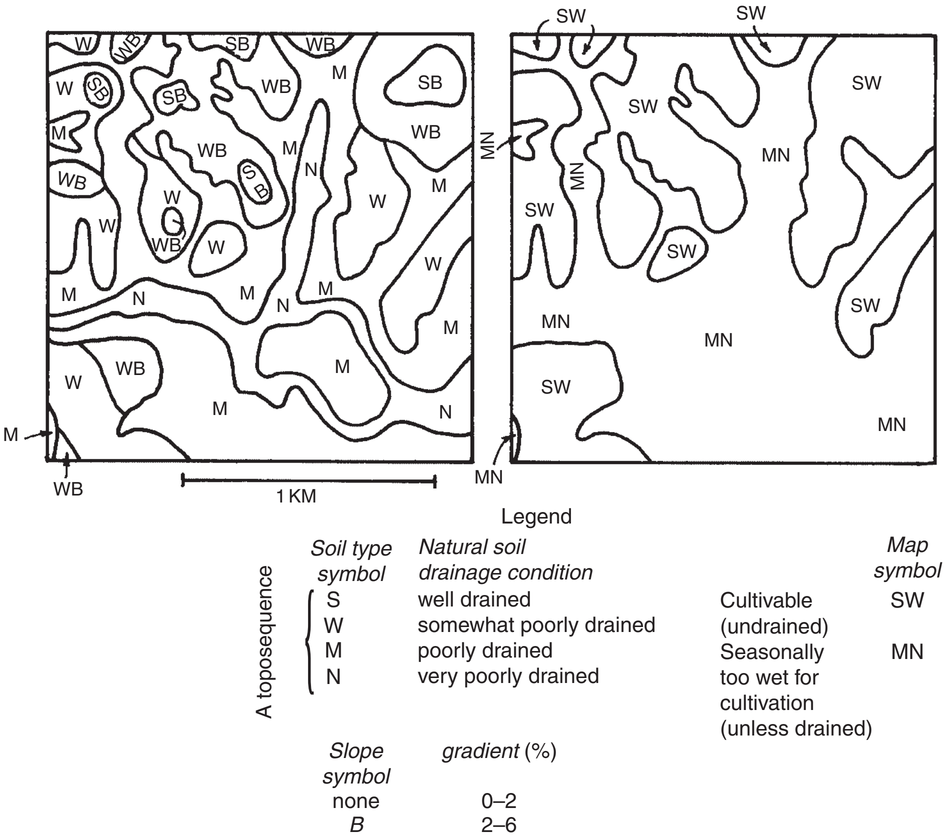
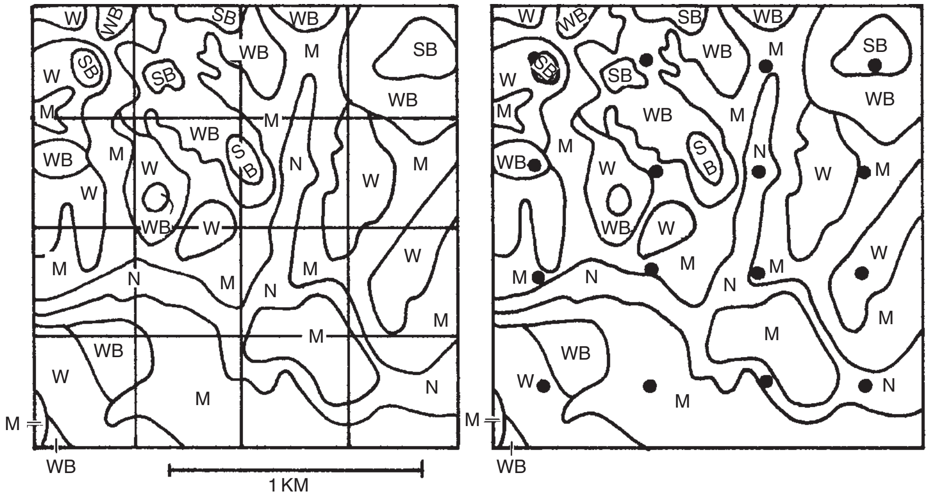


Figure 20.5. Spatial generation of a soil map showing bodies of a four-member toposequence. (*Left*) Detailed soil map showing slope phase delineations. (*Right*) Partly spatially generalized soil map that separates those areas needing drainage from those not needing it. Proportionate extents of soils in this soilscape are S = 12%, W = 26%, M = 43%, N = 19%. (Map based on part of map sheet no. 9, Bartelme 1977)

soil properties identified by each of the map units combined (Figure 20.5). This strategy requires significant changes in the original legend of the primary source map used for the generalization to indicate the composition of the combined units resulting from the generalization and the likelihood of great taxonomic dissimilarities of the combined units.

Naming and describing spatially generalized map units can take three forms (Fridland 1976b). These include: (1) naming the unit according to the predominant soil where one soil is clearly dominant; (2) generalization in which the membership of the combined unit is indicated by soil names but without indicating dominance by one and without indications of the relative proportions or percentages occupied by each soil in the unit; and (3) generalization in which the percentages or relative proportions of each named soil component are specified.



W	WB	M	WB
W	WB	M	W
M	M	M	M
W	M	M	M

SB	WB	M	SB
WB	M	N	M
M	M	M	W
W	M	N	N

	Proportionate areas	
Unit	By Quadrat	By Dot
	%	%
SB	–	12
W	25	13
WB	19	13
M	56	43
N	–	19
	100	100

Figure 20.6. Two examples of sampling generalization of a detailed soil map, using cells (quadrats) (*left*) and dots (*right*). (Map based on part of map sheet no. 9, Bartelme 1977)

Sampling Generalization. This strategy involves giving a single soil name to a map unit based on sampling the taxonomic entities present within a sampling area on the base map. This is a form of spatial generalization that either denotes the soil present at a point where a randomly placed point falls within a sampling area such as a pixel or grid cell or that denotes the soil unit that occupies the largest area within the sampling cell or pixel (Figure 20.6). It should be noted that those soils occurring in long, narrow delineations or other geometric shapes less likely to be selected are apt to be underrepresented in this process. This generalization technique has been widely used to incorporate polygon-based soil survey maps into raster-based GIS applications.

Of these four strategies, spatial generalization by sampling is most frequently used by mapmakers despite the fact that it develops the most taxonomically diverse units. Soil scientists often use a combination of graphic and taxonomic generalization to produce small-scale state, regional, and national maps.

Principles of Soil Map Generalization. We conclude this discussion of soil map generalization procedures and strategies with these principles (Hole and Campbell 1985):

1. The overall quality of a generalized soil map depends on the kind of landscape mapped and the generalization processes applied to the detailed primary source map.
2. A generalized map can be as rigorous and precise as a detailed soil survey map in that it should have known properties, including estimates of error assigned to map units and locations on the generalized map.
3. Users of generalized maps should be informed of strong points, deficiencies, and errors of each map unit comprising the map.
4. Some soil situations and occurrences do not lend themselves to convenient generalization into homogeneous map units.
5. The basic problem in map generalization is that of partitioning diversity into broad categories, subject to the spatial constraints of size, shape, and arrangement of map units in the detailed primary source map.

Figures 20.7, 20.8, 20.9, and 20.10 illustrate how an area as represented on a 1:15,840 scale soil map (Figure 20.7) is progressively represented at smaller and smaller scales. Table 20.3 contains an estimate of the global extent of order and suborder taxa synthesized from generalized and schematic maps.

Assessing Map Unit Composition

Users of soil surveys are pressing for quantification of soil variability and statistical data on specific soil properties and soil performance within map units (Miller 1978; Arnold and Wilding 1991). Most concern is related to statistical evaluations of how well first and second order soil maps produced by "free survey" techniques represent the spatial distribution of pedons in the soilscape.

In selecting a method of assessing spatial variability, we must balance the objectives and results with (or against) desired scale factors (size of tract), statistical

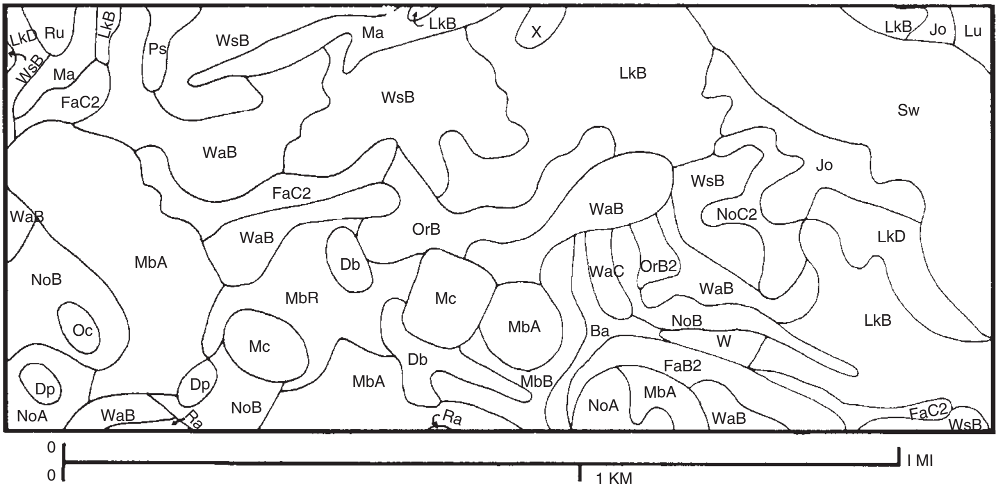


Figure 20.7 Example of a standard soil survey map, cartographically detailed, categorically detailed. (After Horton 1967; Soil Survey Division 2010)

Legend

Ba = Bibb soils, local alluvium. Coarse-loamy, siliceous, active, acid, thermic Typic Fluvaquents.
Db = Dunbar fine sandy loam. Fine, kaolinitic, thermic Aeric Paleaquults.
Dp = Duplin sandy loam. Fine, kaolinitic, thermic Aquic Paleudults.
FaB2 = Faceville loamy sand, 2–6% slopes, eroded. Fine, kaolinitic, thermic Typic Kandiudults.
FaC2 = Faceville loamy sand, 6–10% slopes, eroded. Fine, kaolinitic, thermic Typic Kandiudults.
Jo = Johns loamy sand. Fine-loamy over sandy or sandy skeletal, siliceous, semiactive, thermic Aquic Hapludults.
LkB = Lakeland sand, 0–10% slopes. Thermic, coated Typic Quartzipsamments.
LkD = Lakeland sand, 10–20% slopes. Thermic, coated Typic Quartzipsamments.
Lu = Lumbee loamy sand. Fine-loamy over sandy or sandy skeletal, siliceous, subactive, thermic Typic Endoaquults.
Ma = Mantachie soils, local alluvium. Fine-loamy, siliceous, active, acid, thermic Fluventic Endoaquepts.
MbA = Marlboro loamy sand, 0–2% slopes. Fine, kaolinitic, thermic Typic Paleudults.
Mc = McColl loam. Fine, kaolinitic, thermic Typic Fragiaquults.
NoA = Norfolk loamy sand, 0–2% slopes. Fine-loamy, kaolinitic, thermic Typic Kandiudults.
NoB = Norfolk loamy sand, 2–6% slopes. Fine-loamy, kaolinitic, thermic Typic Kandiudults.
NoC2 = Norfolk loamy sand, 6–10% slopes, eroded. Fine-loamy, kaolinitic, thermic Typic Kandiudults.
Oc = Ocilla loamy sand. Loamy, siliceous, semiactive, thermic Aquic Arenic Paleudults.
OrB = Orangeburg loamy sand, 2–6% slopes. Fine-loamy, kaolinitic, thermic Typic Kandiudults.
OrB2 = Orangeburg loamy sand, 2–6% slopes, eroded. Fine-loamy, kaolinitic, thermic Typic Kandiudults.
Ps = Plummer sand. Loamy, siliceous, subactive, thermic Grossarenic Paleaquults.
Ra = Rains fine sandy loam. Fine-loamy, siliceous, semiactive, thermic Typic Paleaquults.
Ru = Rutlege loamy sand. Sandy, siliceous, thermic Typic Humaquepts.
Sw = Swamp.
WaB = Wagram loamy sand, 2–6% slopes. Loamy, kaolinitic, thermic Arenic Kandiudults.
WaC = Wagram loamy sand, 6–10% slopes. Loamy, kaolinitic, thermic Arenic Kandiudults.
WsC = Wagram sand, thick surface, 0–6% slopes. Loamy, kaolinitic, thermic Arenic Kandiudults.
x = Gravel pit.
w = Water.

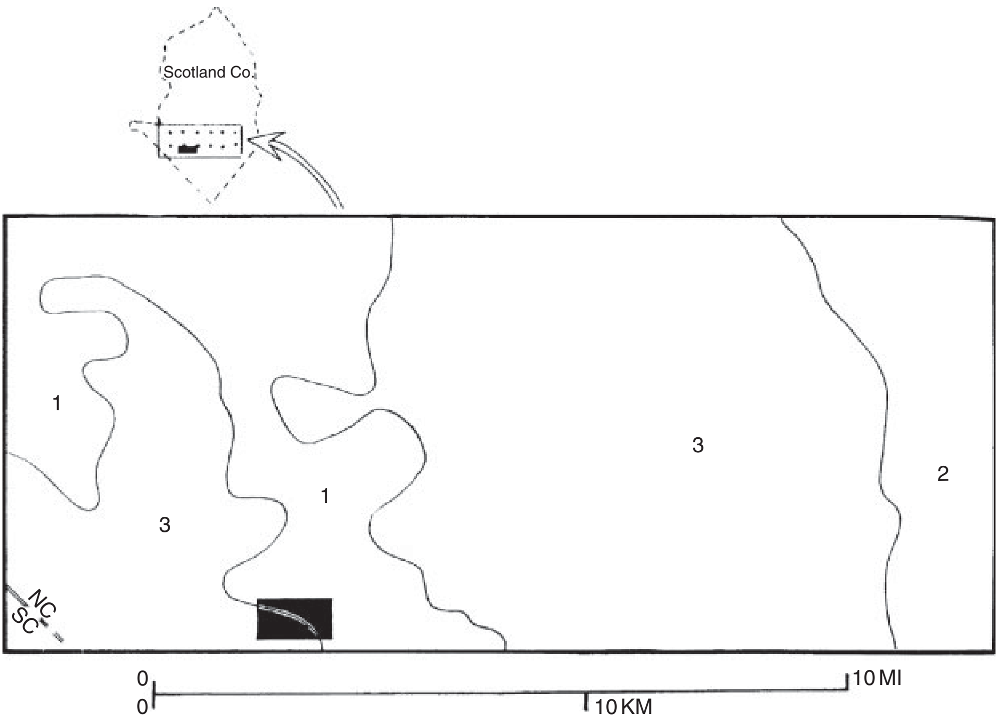


Figure 20.8. Example of a soil map cartographically generalized and categorically detailed. (After Horton 1967.) The dark rectangle represents the area shown in Figure 20.7.

Legend

1 = Lakeland-Gilead. Nearly level to moderately steep sandy soils of the Sandhills.
2 = Eustis-Wagram-Kenansville. Nearly level to sloping, well-drained, or somewhat excessively drained sands and loamy sands on broad ridges.
3 = Marlboro-Norfolk-McColl. Nearly level to sloping, well-drained soils on broad upland ridges, and wet soils in oval-shaped Carolina bays.

efficiency and soundness, prediction precision and accuracy needed, and the cost and time required for sampling and analyzing. Consideration must also be given to which method is most likely to continue to provide useful data for a tract as our technology and capability for assessing spatial variability improves.

Taxonomic Purity. Purity of a soil map unit often is assessed as degree of homogeneity with respect to the taxonomic unit for which the map unit is named. A taxonomically pure map unit would consist entirely of pedons having characteristics that are in the range permitted by the definition of a specific taxon. Such an occurrence is unlikely except at the largest map scales and broadest taxonomic categories. The taxonomic purity of map unit delineations commonly shown on older detailed soil maps produced by the U.S. National Cooperative Soil Survey and using series names varies,

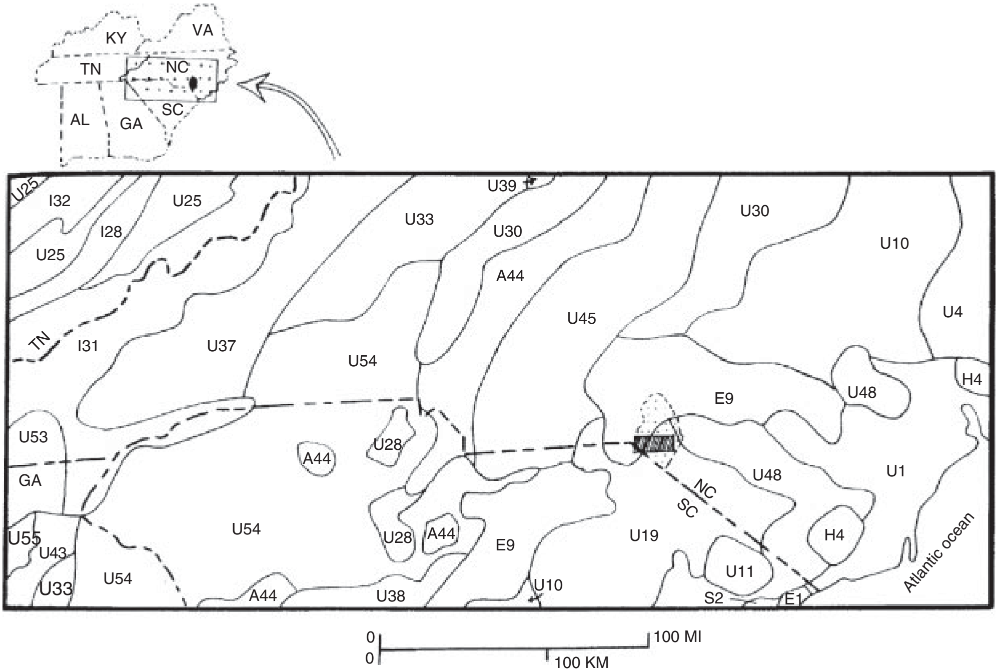


Figure 20.9. Example of a soil map cartographically much generalized. (After Buol 1973.) Line-shaded rectangle represents the area shown in Figure 20.7.

Legend

A44 = Hapludalfs, Kandiudults, Rhodudults
E1 = Sulfaquents
E9 = Quartzipsamments, Kandiudults, Fragiudults
H4 = Medisaprists, Humaquepts
I28 = Eutrudepts, Hapludalfs
I31 = Dystrudepts, Hapludults
I32 = Dystrudepts, Hapludults, Rockland
S2 = Haplaquods, Humaquepts, Quartzipsamments
U1 = Endoaquults, Endoaquods, Psammaquents
U4 = Endoaquults, Hapludults
U10 = Paleudults, Kandiudults (rolling)
U11 = Paleudults, Kandiudults, Quartzipsamments, Psammaquents
U19 = Paleudults, Kandiudults, Paleaquults
U25 = Paleudults, Kandiudults, Hapludults
U28 = Rhodudults, Hapludalfs, Paleudalfs
U30 = Hapludults, Kanhapludults (thermic)
U33 = Hapludults, Kanhapludults, Dystrudepts (hilly, thermic)
U37 = Hapludults, Kanhapludults, Dystrudepts (steep, mesic)
U38 = Hapludults, Kanhapludults, Dystrudepts (steep, thermic)
U39 = Hapludults, Kanhapludults, Dystrudepts (undulating, thermic)
U43 = Hapludults, Kanhapludults, Dystrudepts, Rhodudults
U45 = Hapludults, Kanhapludults, Hapludalfs
U48 = Hapludults, Kanhapludults, Endoaquults, Psammaquents
U53 = Hapludults, Kanhapludults, Rhodudults, Dystrudepts
U54 = Hapludults, Kanhapludults, Rhodudults,

but studies suggest that the figure may be about 65% (Powell and Springer 1965; Wilding et al. 1965). Beckett and Webster (1971) report in a review paper that in terms of taxonomic units the percent "purity" of soil map units ranges from about 50% for series and types to 75% for soil orders.

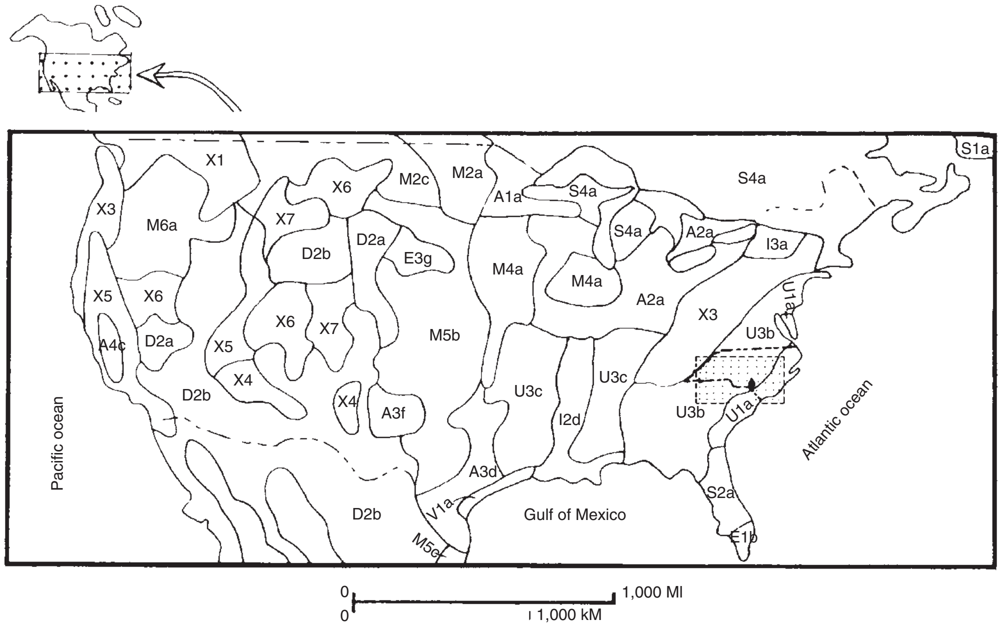


Figure 20.10. Example of a soil map very much generalized cartographically and categorically. The stippled rectangle represents the area shown in Figure 20.9.

Legend

A1a = Udalfs: cool; with Histosols; frigid and cryic soil temperature regimes common.
A2a = Udalfs: temperate to hot, usually moist; with Aqualfs.
A3d = Ustalfs: temperate to hot, dry more than 90 cumulative days during periods when temperature is suitable for plant growth; with Usterts.
A3f = Ustalfs with Ustolls.
A4c = Xeralfs: temperate or warm, moist in winter and dry more than 60 consecutive days in summer; with Xerults.
D2a = Argids: with horizons of clay accumulation; with Fluvents.
D2b = Argids with Torriorthents.
E1b = Aquents: seasonally or perennially wet; Psammaquents with Haplaquents.
E3g = Psamments: sand or loamy sand textures; Ustipsamments with Ustolls.
I2d = Aquepts: seasonally wet.
I3a = Udepts: thin, light-colored surface horizons and little organic matter; Dystrudepts with Fragiudepts.
M2a = Udolls: cool or cold; with Aquolls.
M2c = Ustolls with Torriorthents.
M4a = Udolls: temperate or warm, usually moist; with Aquolls.
M5b = Ustolls: temperate to hot, dry more than 90 cumulative days in the year; with Ustalfs.
M5c = Ustolls with Usterts.
M6a = Xerolls: cool to warm, moist in winter and dry more than 60 consecutive days in summer; with Xerorthents.
S1a = Spodosols with Cryalfs; cryic soil temperature regime.
S2a = Aquods: seasonally wet; Haplaquods with Quartzipsamments.
S4a = Orthods: with accumulation of organic matter, iron, and aluminum in subsurface horizons; Haplorthods with Udalfs.
U1a = Aquults: seasonally wet; Endoaquults with Udults.
U3b = Kanhapludults, Kandiudults with Udalfs.
U3c = Udults with Udalfs.
V1a = Uderts: usually moist in some part in most years, cracks open fewer than 90 cumulative days in the year; with Uderts.
X1 = Cryepts, Cryods, and Cryorthents in mountainous areas.
X3 = Udalfs, Udults, Udepts, and Udorthents in mountainous areas.
X4 = Ustalfs, Ustepts, Ustolls, and Ustorthents in mountainous areas.
X5 = Xeralfs, Xerepts, Xerolls, and Xerorthents in mountainous areas.
X6 = Torric great groups of Entisols; Aridisols in mountainous areas.
X7 = Alfisols, Entisols, Inceptisols, and Mollisols in mountainous areas with ustic to udic soil moisture regimes and mesic to cryic soil temperature regimes.

Table 20.3. Global distribution of orders and suborders and by continent. Column totals may be slightly different than component totals due to rounding errors. (Adapted from World Soil Resources, USDA-NRCS data, unpublished 2010)

		Global		Africa		Asia	
Soil Order	Suborder	10^3 km^2	%	10^3 km^2	%	10^3 km^2	%
Gelisols	Histels	1,006	0.77	–[a]	–	478	1.09
	Turbels	5,065	3.88	–	–	1,973	4.49
	Orthels	5,692	4.36	–	–	4,372	9.96
		11,764	**9.00**	**–**	**–**	**6,823**	**15.54**
Histosols	Folists	–	–	–	–	–	–
	Fibrists	196	0.15	–	–	136	0.31
	Hemists	981	0.75	–	–	275	0.63
	Saprists	331	0.25	17	0.06	249	0.57
		1,507	**1.15**	**17**	**0.06**	**660**	**1.50**
Spodosols	Aquods	167	0.13	–	–	13	0.03
	Cryods	2,574	1.97	–	–	65	0.15
	Humods	57	0.04	–	–	19	0.04
	Orthods	649	0.50	–	–	13	0.03
	Gelods	1,111	0.85	–	–	107	0.24
		4,558	**3.49**	**–**	**–**	**217**	**0.50**
Andisols	Cryands	251	0.19	–	–	<57	0.13
	Torrands	1	0.00	1	0.00	<1	0.00
	Xerands	32	0.02	–	–	–	–
	Vitrands	281	0.21	1	0.00	20	0.04
	Ustands	62	0.05	12	0.04	19	0.04
	Udands	275	0.21	33	0.11	109	0.25
	Gelands	62	0.05	–	–	59	0.13
		963	**0.74**	**46**	**0.16**	**263**	**0.60**
Oxisols	Aquox	322	0.25	321	1.08	–	–
	Torrox	31	0.02	9	0.03	–	–
	Ustox	3,115	2.38	1,712	5.75	19	0.04
	Perox	1,167	0.89	89	0.30	55	0.12
	Udox	5,233	4.01	1,744	5.85	77	0.18
		9,868	**7.55**	**3,875**	**13.01**	**151**	**0.34**
Vertisols	Aquerts	5	0.00	<1	0.00	3	0.01
	Cryerts	17	0.01	–	–	–	–
	Xererts	98	0.08	12	0.04	47	0.11
	Torrerts	894	0.68	194	0.65	65	0.15
	Usterts	1,770	1.35	716	2.40	599	1.36
	Uderts	383	0.29	66	0.22	110	0.25
		3,167	**2.42**	**989**	**3.32**	**824**	**1.88**
Ardisols	Cryids	1,036	0.79	<1	0.00	400	0.91
	Salids	1,287	0.98	162	0.55	995	2.27
	Gypsids	680	0.52	360	1.21	318	0.73
	Argids	4,678	3.58	349	1.17	1,107	2.52
	Calcids	4,887	3.74	1,806	6.06	2,184	4.97
	Cambids	2,919	2.23	848	2.85	1,106	2.52
		15,487	**11.85**	**3,525**	**11.83**	**6,111**	**13.92**

Australia/Oceania		Europe		South America		North America	
10^3 km^2	%	10^3 km^2	%	10^3 km^2	%	10^3 km^2	%
–	–	77	0.80	–	–	451	2.08
–	–	54	0.56	23	0.13	3,015	13.91
–	–	208	2.17	55	0.31	1,056	4.87
–	**–**	**339**	**3.54**	**79**	**0.44**	**4,523**	**20.87**
–	–	–	–	–	–	–	–
–	–	–	–	–	–	59	0.27
1	0.01	205	2.14	10	0.06	489	2.26
–	–	<1	0.00	25	0.14	39	0.18
1	**0.01**	**206**	**2**	**36**	**0.20**	**588**	**2.71**
–	–	99	1.04	–	–	55	0.25
<1	0.00	1,013	10.56	15	0.08	1,482	6.84
33	0.41	4	0.05	<1	0.00	<1	0.00
29	0.36	265	2.76	<1	0.00	341	1.58
–	–	402	4.19	–	–	601	2.77
61	**0.77**	**1,784**	**18.60**	**16**	**0.09**	**2,480**	**11.44**
<1	0.00	25	0.27	72	0.41	96	0.44
–	–	–	–	<1	0.00	<1	0.00
–	–	13	0.13	2	0.01	17	0.08
29	0.37	<1	0.00	49	0.28	181	0.84
1	0.01	2	0.02	11	0.06	18	0.08
32	0.40	<1	0.01	82	0.46	19	0.09
–	–	<1	0.00	2	0.01	<1	0.00
62	**0.78**	**41**	**0.43**	**218**	**1.23**	**332**	**1.53**
<1	0.01	–	–	–	–	–	–
4	0.05	–	–	16	0.09	2	0.01
54	0.67	–	–	1,323	7.51	1	0.01
9	0.11	–	–	1,013	5.73	<1	0.00
25	0.31	–	–	3,383	19.12	4	0.02
92	**1.15**	**–**	**–**	**5,741**	**32.46**	**8**	**0.04**
–	–	–	–	1	0.01	–	–
–	–	17	0.18	<1	0.00	<1	0.00
20	0.25	17	0.18	1	0.01	1	0.00
586	7.33	–	–	6	0.04	42	0.20
243	3.03	28	0.29	39	0.22	145	0.67
14	0.18	20	0.21	106	0.60	66	0.31
863	**10.79**	**82**	**0.86**	**153**	**0.87**	**255**	**1.18**
–	–	<1	0.00	160	0.90	476	2.19
78	0.98	8	0.08	33	0.18	10	0.05
–	–	2	0.02	–	–	–	–
1,591	19.88	64	0.67	502	2.84	1,065	4.91
599	7.49	1	0.01	103	0.58	194	0.90
337	4.21	29	0.30	431	2.44	169	0.78
2,604	**32.56**	**104**	**1.09**	**1,228**	**6.94**	**1,914**	**8.83**

Table 20.3. *Concluded.*

Soil Order	Suborder	Global 10³ km²	Global %	Africa 10³ km²	Africa %	Asia 10³ km²	Asia %
Ultisols	Aquults	1,285	0.98	56	0.19	257	0.59
	Humults	380	0.29	36	0.12	178	0.40
	Udults	5,540	4.24	603	2.02	3,066	6.98
	Ustults	3,340	2.56	1,356	4.55	718	1.63
	Xerults	19	0.01	<1	0.00	3	0.01
		10,563	**8.08**	**2,050**	**6.88**	**4,221**	**9.61**
Mollisols	Albolls	7	0.01	–	–	7	0.02
	Aquolls	118	0.09	–	–	35	0.08
	Rendolls	261	0.20	<1	0.00	63	0.14
	Xerolls	923	0.71	75	0.25	256	0.58
	Cryolls	2,464	1.89	–	–	1,593	3.63
	Ustolls	3,937	3.01	4	0.01	949	2.16
	Udolls	1,263	0.97	–	–	131	0.30
	Gelolls	156	0.12	–	–	151	0.34
		9,128	**6.99**	**80**	**0.27**	**3,184**	**7.25**
Alfisols	Aqualfs	1,029	0.79	393	1.32	35	0.08
	Cryalfs	2,531	1.94	–	–	729	1.66
	Ustalfs	6,024	4.61	2,647	8.89	1,102	2.51
	Xeralfs	893	0.68	84	0.28	214	0.49
	Udalfs	2,678	2.05	165	0.55	271	0.62
		13,156	**10.07**	**3,289**	**11.04**	**2,351**	**5.35**
Inceptisols	Aquepts	3,657	2.80	369	1.24	1,705	3.88
	Anthrepts	450	0.34	184	0.62	29	0.07
	Cryepts	2,598	1.99	<1	0.00	1,495	3.41
	Ustepts	2,230	1.71	764	2.56	589	1.34
	Xerepts	683	0.52	167	0.56	177	0.40
	Udepts	4,102	3.14	219	0.73	1,347	3.07
	Gelepts	6,043	4.63	–	–	4,752	10.82
		19,764	**15.13**	**1,703**	**5.72**	**10,094**	**22.99**
Entisols	Aquents	109	0.08	43	0.14	65	0.15
	Psamments	4,447	3.40	2,271	7.62	122	0.28
	Fluvents	3,056	2.34	666	2.24	1,284	2.93
	Orthents	15,834	12.12	7,896	26.50	4,634	10.55
		23,446	**17.94**	**10,876**	**36.51**	**6,105**	**13.90**
Miscellaneous	Salt	145	0.11	35	0.12	52	0.12
	Shifting sands	5,341	4.09	3,001	10.07	1,720	3.92
	Rock	1,803	1.38	306	1.03	1,134	2.58
	Ice	2,210		–		257	
			5.58	3,342	11.22	3,163	6.62
Total		**130,658**	**100.00**	**29,792**	**100.00**	**43,911**	**100.00**

[a]none reported.

Australia/Oceania		Europe		South America		North America	
10^3 km^2	%	10^3 km^2	%	10^3 km^2	%	10^3 km^2	%
76	0.95	–	–	692	3.91	205	0.94
25	0.32	–	–	–	–	142	0.65
88	1.10	4	0.04	858	4.85	921	4.25
71	0.89	–	–	1,058	5.98	138	0.64
–	–	–	–	<1	0.00	15	0.07
260	**3.25**	**4**	**0.04**	**2,608**	**14.74**	**1,420**	**6.55**
–	–	–	–	–	–	–	–
–	–	15	0.16	–	–	68	0.31
12	0.15	127	1.33	–	–	59	0.27
58	0.73	372	3.88	<1	0.00	162	0.75
–	–	153	1.59	51	0.29	667	3.08
22	0.27	899	9.37	408	2.31	1,656	7.64
25	0.32	78	0.82	527	2.98	500	2.31
–	–	<1	0.00	1	0.01	3	0.02
117	**1.46**	**1,644**	**17.14**	**989**	**5.59**	**3,115**	**14.37**
126	1.58	374	3.90	56	0.32	45	0.21
<1	0.00	1,162	12.11	2	0.01	639	2.95
422	5.28	390	4.07	1,028	5.81	435	2.01
267	3.34	129	1.35	28	0.16	170	0.79
228	2.85	502	5.23	675	3.82	837	3.86
1,044	**13.05**	**2,556**	**26.66**	**1,789**	**10.12**	**2,126**	**9.81**
3	0.03	195	2.03	414	2.34	971	4.48
–	–	–	–	227	1.28	10	0.05
9	0.11	351	3.66	261	1.47	482	2.22
15	0.18	370	3.85	181	1.02	313	1.44
<1	0.00	321	3.35	8	0.05	9	0.04
252	3.15	768	8.01	934	5.28	583	2.69
–	–	99	1.03	2	0.01	1,191	5.49
279	**3.49**	**2,102**	**21.92**	**2,026**	**11**	**3,559**	**16.42**
–	–	–	–	1	0.01	–	–
1,251	15.64	22	0.23	781	4.42	–	–
9	0.11	343	3.57	631	3.57	123	0.57
762	9.52	299	3.12	1,122	6.35	1,121	5.17
2,022	**25.28**	**664**	**6.92**	**2,536**	**14.33**	**1,243**	**5.74**
–		–		49	0.27	10	0.04
594	7.43	22	0.23	4	0.02	–	–
–	–	41	0.43	219	1.24	102	0.47
–		89		14		1,849	
594	7.43	152	0.66	285	1.53	1,961	0.52
8,000	**100.00**	**9,590**	**100.00**	**17,690**	**100.00**	**21,676**	**100.00**

The practical value of determining the taxonomic purity of soil map units is questionable. Many taxonomic units are equally suited for several potential land uses. A low degree of taxonomic purity does not indicate a low degree of interpretive purity. Several comparisons have been reported, and specific values differ slightly, but in most studies, map units with taxonomic purity of well less than 50% are found to have interpretive purity of 80% or higher (Young et al. 1997; Wilding et al. 1965; McCormack and Wilding 1969; Nordt et al. 1991).

Interpretive Purity. The practical land manager focuses attention on the degree to which inclusions of impurities within map unit delineations influence his intended land use, and the feasibility of overcoming obstacles to management operations that such inclusions may present (Oschwald 1966; Riecken 1963). Present practice in the U.S. National Cooperative Soil Survey is that of consociations, complexes, associations, or undifferentiated map units that assess purity based on the area of included soils that are similar and dissimilar for major interpretations of the map unit delineations (Soil Survey Division Staff 1993). This provides most users of soil maps with a realistic and practical assessment they can use to interpret the suitability of areas of land delineated in map units for their intended use. Not all potential users, especially those who seek to use small parcels of land, will find their needs satisfied because of map-scale limitations (Table 20.1) endemic to most published soil surveys, and more detailed, larger-scale surveys are required.

Quantifying Spatial Variability in Map Units. Spatial variability among pedons in a soilscape is natural. It is only possible to reduce the variability and the inclusions in map units by conducting detailed, large scale-soil mapping at relatively high cost. Quantification of similar and dissimilar inclusions in map units adds confidence to the soil map, provides information to our understanding of soil formation, and improves utility of soil maps for various land-use interpretations. Wilding and Drees (1983) list several kinds of information that studies of spatial variability of soils will provide, including the following:

1. Estimations of central tendencies via means, etc. and variance statistics for specific soil classes.
2. Quantification for soil genesis studies of both the effects of pedogenic processes and external soil-forming factors.
3. More quantitative data on the composition of soil map units.
4. Better sampling designs and statistical modes for soil survey and soil-formation study purposes.

To these we would add that quantification of spatial variation in existing soil surveys serves to better identify those map units where the cost of more detailed surveys is needed for particular land use interpretations is justified. Furthermore, measures of soil variability will facilitate efforts to develop more quantitative databases using existing soil surveys.

Numerous techniques have been devised or adapted from geological mapping, mineral prospecting, and other disciplines that quantitatively characterize populations of natural objects. These include random, grid, and transect sampling, and geostatistics.

Random Point Sampling. This procedure for determining map unit composition involves randomly selecting delineations or sites within delineations. Map units may or may not be stratified for such features as slope, erosion, areal extent, and/or other features as dictated by the objectives of the study. The random selections should be done in the office on existing maps so the selection by observing field situations is not biased (Wilding and Drees 1983). This technique has been frequently and widely used in soil fertility, soil management, and soil physics field work, but has been employed only in a very limited way in studies of soil map unit and soilscape variability (Wilding et al. 1965; McCormack and Wilding 1969). This technique has advantages of being statistically sound and unbiased and is relatively simple, efficient, and economical. However, as pointed out by Wilding and Drees (1983), random samplings tend to cluster spatially and are not likely to detect and measure systematic spatial variation. Random locations do not yield maps of uniform variability over the study areas and are not suitable for numerical classification. Therefore, although useful in estimating quality of soil maps, random samples are less useful in depicting and quantifying spatial variability in a manner most useful for predicting suitability of and potential problems with land use of a particular small tract.

Grid Sampling. A grid with suitable spacing is placed on the map to be studied. Sites can be selected at intersections of the grid lines or within the grid cells (Wilding and Drees 1983).

Grid sampling provides equally spaced observations that can readily be used for computerized applications and for certain kinds of statistical analyses (Wilding et al. 1965; Hock et al. 1973; Burgess and Webster 1980a; Wilding and Drees 1983). This technique reveals any systematic variation across the tract under study and provides ease in plotting, especially with use of computers, and in graphic analyses. However, it does not lend itself well to numerical classification or to computerized preparation of isovariograms, is not as preferable statistically as pure random sampling, and requires considerable surveying effort to ensure proper location of the grid points in the field.

Transect Sampling. Soil variability may be determined along transects in one dimension (Fridland 1976b; Hajek 1977; Nwadialo 1978), or in two dimensions (Campbell 1977), using statistical methods described by the workers cited. Results show that in a complex soilscape, a shorter transect will suffice to span the full range of spatial variability than in a simpler soilscape. Three methods of transect sampling,

(1) simple random, (2) cluster, (3) two-stage random, have been used to evaluate the central tendency and variance of soil properties in map units (Young et al. 1998). Each method has merit, but none is universally applicable. Differences and prominence of soil properties across soilscape continuums represented by map units dictate that the direction of each transect fairly traverses the delineation and avoids concentrations in the spatial center of the delineation, although transect data from diagnostic portions of the map unit may be important to reveal the spatial nature of actual soil variability. Transect sampling is frequently used to obtain quantitative estimates of pedon populations within soil map units and set standards for soil mapping.

Geostatistics. The theory of regionalized variables is the basis of geostatistics (Journel and Huijbregts 1978; Burgess and Webster 1980a, 1980b; Webster and Burgess 1980; Wilding and Drees 1983). Any variable that is distributed in space and whose values are related to its location (spatial dependence) is considered regionalized. From this theory is derived the semivariogram and the prediction method called kriging, as means of spatial prediction of soil properties. Soil samples are collected according to statistically designed grid or point procedures, the number being related to the degree of statistical precision required, desired width of the confidence interval, and the coefficient of variation acceptable.

Isovariograms and kriging have been used for study of the spatial variability of a number of soil properties. Heuvelink and Webster (2001) have reviewed the development of kriging as a tool for modeling soil variation. Early examples can be found in the papers by Burgess and Webster (1980a, 1980b), Webster and Burgess (1980), Nielson and Bouma (1985), and Doucette (1983). Taxonomic variation, spatial variability, and taxonomic composition of map units have been studied by Edmonds et al. (1985a, 1985b). The studies reported vary widely in sampling density, procedures used, and in properties selected for analysis. According to a summary by Wilding and Drees (1983), estimates of the mean value with ± 10% at the 95% confidence level are unrealistic for many soil properties because of the large number of profile samples required.

Kriging can be very costly and time-consuming when assessing and quantifying soil properties if applied routinely in large areas in attempts to improve prediction power and the quantitative information of soil surveys. In some cases, kriging may ignore the proven relationships between soil properties and terrain attributes. In addition, kriging is unable to handle abrupt boundaries that may exist in nature, such as changes in lithology and topographic breaks (Heuvelink and Webster 2001). In some cases, prediction of soils and their properties can be greatly enhanced by using secondary variables that are correlated with the primary variable to be predicted, a technique known as co-kriging (McBratney et al. 2003). Co-kriging commonly utilizes terrain attributes derived from digital elevation models or remote sensing as secondary variables; these techniques have the advantage of being able to provide large, detailed datasets (McBratney et al. 2003).

Perhaps a minor point but nevertheless troubling fact of life related to all observations of soil via soil auger or mechanical probes is the presence of rodent burrows, abandoned fence-post holes, and decayed tree stump holes. Such observations are not representative of pedons and are best discarded as inconsequential to map unit purity investigations.

Relation Between Soilscape and Soil Classification

Information about soilscapes and their component polypedons forms a bridge between the real world and the abstractions of taxonomic units and soil classification systems. It should be clearly understood by soil scientists that the arrangement of soils with respect to each other in a classification system might bear little resemblance to their geographic arrangement within a soilscape. This is only reasonable, because the geographic arrangement of soils is not always the same from soilscape to soilscape, whereas soil classification systems must provide for placement of all soils using only criteria observable within, but not outside of the pedon. Soilscape identification, that is, soil mapping, must account for the natural arrangement of soil bodies in the real world. More homogeneous soil groupings can often be obtained by landform identification within a given area than by taxonomic identification (Young and Hammer 2000). Young and Hammer point out that taxonomic classes are "bound-from-without" by divisions inappropriate to existing soil patterns in many landscapes rather than "bound-from-within" by processes controlled by soil-forming factors within landscapes.

An illustration of nonconformity between taxonomic and soilscape placement of soil is given in Figure 20.11. In that figure, a soilscape of Miami silt loams (Oxyaquic Hapludalfs) lies in a "crest" or summit position on a narrow drumlin ridge, 10 to 15 meters above associated Haplaquolls in Dodge County, southeastern Wisconsin. In Wayne County, Indiana, on the other hand, Haplaquolls are in a slight depression in a broad upland above soilscapes of Miami silt loam soils that are on adjacent gentle valley slopes. The Miami pedons at both locations are within the taxonomically defined range of Miami soil characteristics, but each site is likely to have pedons representing a different portion of the taxonomically defined range. Therefore, soil property groupings tend be more precise within any given map unit than within taxonomic groupings, but such map unit groupings of soil characteristics are seldom applicable to other areas and soilscapes.

The size, shape, and spatial association of soilscapes (map unit delineations) have a marked influence on many uses of taxonomically identified kinds of soil, especially in mechanized commercial farming where management of soils for agricultural purposes is carried out in accordance with soil body and soilscape properties. Productivity of soilscapes varies with the character and areal distribution of component soils. Although the pedon characteristics of two soils may be identical, their value for almost any use depends on their areal extent and may drastically change depending upon the size, shape, and spatial association with other soilscapes. Soil map unit delineations serve to communicate both internal pedon properties via taxonomic identification and the spatial dimensions of soils as natural bodies on the earth's surface.

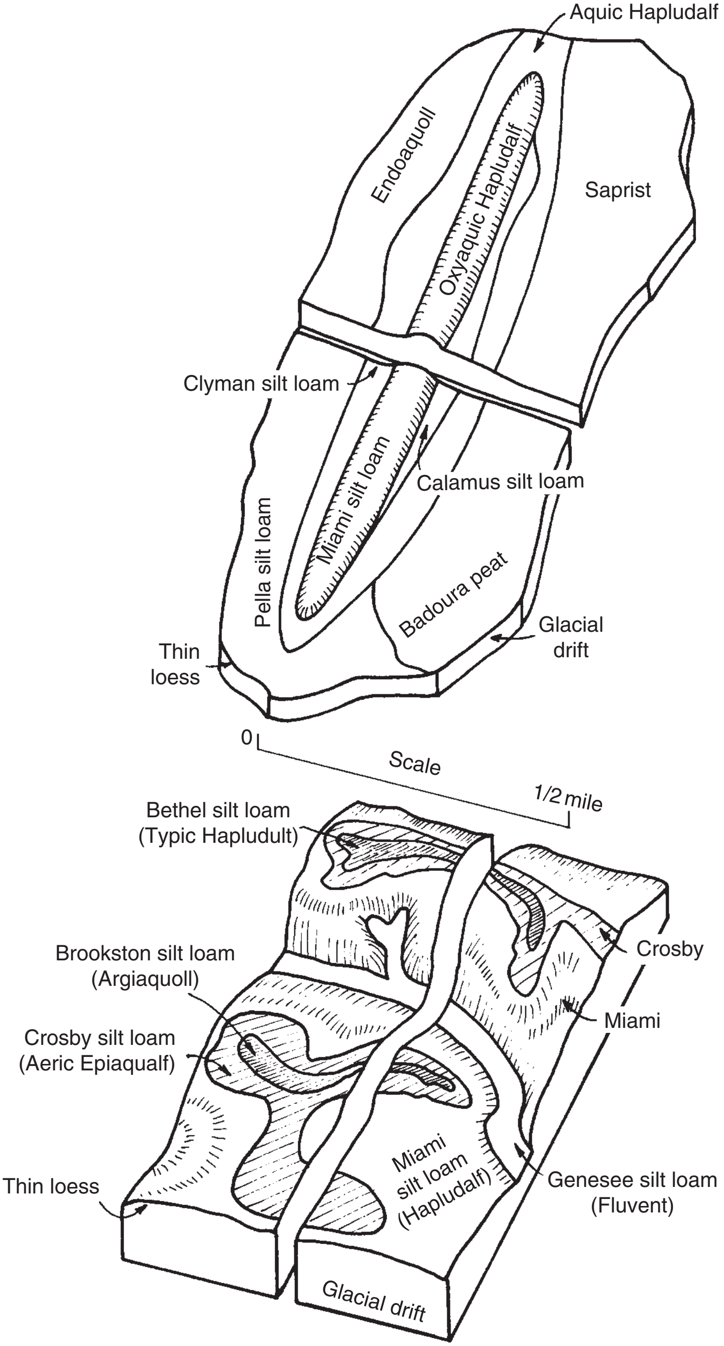


Figure 20.11. Contrasting landscape positions of Miami silt loam, an Oxyaquic Hapludalf, in southern Wisconsin (Hole et al. 1953) and east central Indiana (Bushnell et al. 1930). The *upper* diagram is of a drumlin about three–quarters of a mile long on which the Miami soil occurs, with another Oxyaquic Hapludalf, the moderately well-drained Calamus, on a footslope and surrounded by wetter soils. The *lower* diagram shows the Miami soil on marked slopes (exaggerated in the sketch) with wetter soils on the upland flats.

Perspective

The basic distinction between soil map units and soil taxonomic units is that the latter are abstract concepts that group soils according to specific ranges of soil properties for purposes of scientific categorization, whereas soil map units are cartographic representations of soils as they spatially occur on the earth's surface. Techniques of generalizing large-scale maps are presented, but the natural spatial variability of the soils and how that variability is presented in the most detailed soil maps is a major concern. The technique that best evaluates the natural distribution of soils and conveys that information using map units at this time is best described as a keen observer with knowledge of soil genesis and experience. Although efficient and effective, this technique can be strengthened with quantitative rigor of statistical evaluation and application of newer technologies. Quantifying spatial variability of soil within map unit delineations seeks to improve the predictions made about land. Evaluations of map unit purity have to address whether that evaluation is for taxonomic purity or interpretive purity related to specific soil use practices. Procedures that can be used to quantify map unit variation are geostatistics, random point sampling, grid sampling, and transect sampling. Unfortunately, techniques used to quantitatively evaluate and predict soil properties have limitations, not the least of which is the large amount of data and cost required. As demand increases for spatially explicit, quantitative soil data, one of the challenges facing those working in soil survey will be to further develop ways by which existing soil maps can be combined with newer geocomputational technologies to provide such information. A variety of modeling techniques, geostatistics, fuzzy logic theory, expert knowledge, remote and proximal sensing, and GIS technologies offer a means for doing this.

Interpretations of Soil Surveys and Technical Soil Classification

21

Interpretations of soil surveys are the "proof of the pudding" as the theories, hypotheses, and knowledge expressed in taxonomic systems are put to the test through applied, practical uses. The best test of our principles and technologies of soil genesis, classification, and surveys is how well they can be applied in the generalization of useful, easily understood descriptions of soil qualities and behavior important for various uses. Technical soil classifications are designed for a specific use and purpose, using criteria and characteristics appropriate for that purpose. These criteria and characteristics are generally a restricted subset of the properties used in the natural classification system.

Soil properties can be interpreted in the context of "soil quality." Soil quality is a basic, intrinsic characteristic of a soil that is a reflection of several soil properties and cannot be directly characterized in one measurement, but ordinarily is estimated from a number of different measurements and/or observations. The concept of soil quality continues to evolve, but soil properties and environmental factors that are often included in an assessment of soil quality include soil fertility, potential productivity, resource sustainability, environmental quality, contaminant levels and their effects on use, the environmental cost of agricultural production, and the potential for reclamation of degraded soil (Kellogg 1961; Singer and Ewing 2000).

Soil Survey Interpretations

Soil survey interpretations help predict potentials, limitations, problems, and management needs for soils (Kellogg 1966; Soil Survey Division Staff 1993). Aandahl (1958) defined soil survey interpretation as "the organization and presentation of knowledge about characteristics, qualities and behavior of soils as they are classified and outlined on maps." Soil survey interpretations are prepared to help land users, planners, policy makers and analysts, legislative officials, engineers, and other scientists to transfer technology about the use and management of soils—both agricultural and nonfarm—more accurately. These interpretations are made to accompany or be used with detailed soil surveys and larger-scale maps prepared by the generalization process previously described (Chapter 20), or they are made to be

Soil Genesis and Classification, Sixth Edition. S. W. Buol, R. J. Southard, R. C. Graham and P. A. McDaniel.

used with the schematic or technical single-purpose maps (Soil Survey Staff 1978, 1983a; USDA-NRCS 2009).

Interpretations of detailed soil surveys (those that have not been generalized) have a number of uses (USDA-NRCS 2009): indicating soil potentials for various land uses, pointing out limitations and potential problems with soil qualities (especially for conservation and environmental protection), predicting potential production of soils under various management systems, indicating types of management needed for different soils, and recasting the technical soil information into forms and terminology useful in other scientific disciplines and to land users. The soil information used for these detailed interpretations is contained in published county soil surveys, in digital soil surveys at the NRCS Web Soil Survey site (http://websoilsurvey.nrcs.usda.gov/app/HomePage.htm), or in the USDA-NRCS Soil Survey Geographic (SSURGO) Data Base (http://soildatamart.nrcs.usda.gov). For statistical evaluation of crop response on adjacent mapping units, experimental designs that consider the soil patterns within fields are required (Nelson and Buol 1990).

Interpretations of generalized large-scale maps are usually designed for uses such as regional planning, setting priorities for action programs, depicting regional problems in conservation and environmental protection, estimating productivity on a regional basis, preparing and implementing regional development programs, determining feasibility and potential benefits of proposed irrigation and watershed programs, and assessing regional environmental protection needs and concerns. The soil information used for these regional interpretations is contained in soil maps generalized from county soil survey maps or in the USDA-NRCS State Soil Geographic (STATSGO) Data Base (http://soildatamart.nrcs.usda.gov/USDGSM.aspx).

Interpretations are prepared for taxonomic soil components of soil map units or of the soil map unit as a whole. Most commonly they are prepared for phases of taxonomic units (such as slope and eroded phases) if used with a detailed survey or with maps generalized according to the graphic generalization strategy. If used with taxonomically generalized, schematic, or technical maps, interpretations are generally prepared for mapping units, but may also be prepared from the properties of dominant taxonomic components identified in the map unit. As indicated in Chapter 20, both the preparers and the users of soil interpretations need to be aware of, and cautioned about, the differences between taxonomic units and soil mapping units. This is especially true if taxa with different interpretation requirements are included in a soil map unit. This is particularly important when the interest focuses on small sites such as house lots smaller than 1 ha.

Interpretive Uses of Taxonomic Information and Soil Maps

Some interpretations are for specific soil materials present in soil horizons, but most soil uses involve an area of land. For these uses, the nature and composition of map units, as discussed in Chapter 20, are the basis of soil interpretations. Soil map and taxonomic information is put to a number of agricultural and nonagricultural interpretive uses (Bartelli et al. 1966; Simonson 1974). For agricultural uses these include recommended conservation practices; identification of prime farm land; predictions of erosion losses under various cropping systems; development of productivity ratings of soils; soil potentials and crop yield; land evaluations and tax assessments; soil and farm management recommendations, including those on application of fertilizer; programs for protection of environmental quality from damage by sedimentation and agricultural chemicals; and improvements in agricultural energy conservation. Nonagricultural interpretations of soils are many, and include forest, range, and wildlife management; mined land reclamation; local, state, and regional planning, zoning, and other land use concerns; identification of hydric soils, suitability for locating septic tank filter fields and for municipal sewage effluent and sludge disposal; environmental quality protection with respect to off-farm sediment, dust, and pollution control; local road and highway route location, location of pipelines and power lines; building and real estate development site location; suitability as sources of building materials; valuation of land for tax assessment purposes; and planning the location and layout of outdoor recreation facilities, including parks, campgrounds, paths, and trails.

For these purposes it is more desirable and economical to prepare comprehensive soil maps as a base from which the interpretations can readily be made. If a single-purpose technical classification were to be used as the basis for soil mapping, it is highly probable that such a classification would not be suitable for other types of planning and site evaluation, such as the kinds we have described. Then the mapping and classification would need to be redone at considerable unnecessary expense to satisfy each new purpose. A well-prepared soil map, based on a sound general classification system that has quantitatively defined taxa, is useful over a long period of time as a base for a number of interpretation activities for technical and specific objectives, including those not yet foreseen. Such an approach allows (even forces) consideration and description of the dynamic aspects of each soil and the relating of the soils to various uses.

Technical Soil Maps and Classifications. We distinguish between technical soil maps, classifications, and surveys on the one hand and comprehensive standard soil classification, maps, and surveys on the other. The technical maps and classifications, prepared either by interpretation from the standard maps and classifications or specially and separately prepared, are designed for a specific use and purpose, using criteria and characteristics appropriate for that purpose. The comprehensive standard maps and classifications are general and multipurpose in nature with criteria and characteristics selected according to scientific and taxonomic principles. They are capable of being interpreted for any number of specific, single-purpose objectives and uses.

It is important to understand that there are an almost infinite number of technical systems—one for each particular use of soil. Anyone can make a technical map and classification—either by interpreting existing standard comprehensive scientific surveys and classifications or by a separate soil survey with accompanying technical classification. In order to be useful and sound, such technical systems must, however, use a quantitative nomenclature with class limits based on measurable properties of the soils and/or of the environmental factors that influence soil behavior. The discussion that follows relates to preparation of technical systems interpreted from the comprehensive scientific classification and accompanying standard soil maps—the process that we recommend over special purpose technical soil surveys.

Development of Technical Soil Interpretations. The preparation of interpretations by soil scientists of the National Cooperative Soil Survey ordinarily involves the following four steps (Soil Survey Division Staff 1993): (1) assembling information about the soils and landscapes, (2) modeling other necessary soil characteristics from existing soil data, (3) deriving inferences, rules, and guides for predicting behavior under specific land uses, and (4) integrating these predictions into generalizations for the map unit.

The essential ingredients in the development of soil interpretations are a clear definition of the specific soil use and quantitative expression of the influence of each applicable soil property on the evaluation or interpretation of a soil, or mapping unit, for that specific use. Each technical soil use (of which there is an infinite number) can and should have its own technical classification in order to facilitate communication between soil scientists and the specialists in the particular subject matter area of the soil use. The soil data used to develop the interpretations are contained in the National Soil Information System (NASIS) as described in the *National Soil Survey Handbook* (USDA-NRCS 2009). The NASIS database is managed by the USDA-NRCS and is the repository of the attributes of the taxonomic components of soil map units and the geo-referenced site information and pedon data. The database is populated with actual soil property data from pedon descriptions, laboratory characterization, and map unit characteristics, and with inferred soil properties derived from the actual data. Most of the NASIS data are available via the Soil Data Mart (http://soildatamart.nrcs.usda.gov). Laboratory data and pedon descriptions of sampled and characterized pedons are from a separate database maintained by the National Soil Survey Center in Lincoln, Nebraska (http://soils.usda.gov or http://ssldata.nrcs.usda.gov/).

Because it is not possible for anyone to anticipate all of the possible uses of soils information, the following example of a soil interpretation based on criteria outlined in the *National Soil Survey Handbook* (USDA-NRCS 2009) and on data contained in NASIS, is presented to illustrate the procedure for developing a soil interpretation.

Consider that you are confronted with the question: What makes a good picnic area? In order to address the question clearly, a number of broad criteria must be established first. For example, will only hikers use the picnic area or will vehicles be driven to the area? Are garbage facilities to be provided for on the site, or will the garbage be exported? Will extensive revegetation of the site be required following construction activities? Thus, providing a clear definition of the use must be addressed first. In this example, we will use the following definition of picnic areas: *picnic areas*

Table 21.1. Evaluation of sites for plenic areas

Property	Limits: Slight	Moderate	Severe	Restrictive Feature
Slope (%)	0–8	8–15	>15	Slope
Flooding	None to occassional	Frequent	—	Flood hazard
Depth to high water table (cm)	>75	30–75	<30	Wetness
Fraction 2–75 mm (i.e., wt % of surface layer)	<25	25–50	>50	Small stones
USDA texture[a,c] (surface layer)	—	—	SC, SIC, C	Too clayey
USDA texture (surface layer)	—	LCOS, LS, LFS, VFS	COS, S, FS	Too sandy
USDA texture (surface layer)	—	—	Muck, peat	Excess humus
USDA texture[b] (surface layer)	—	SIL, SI, VFSL, L	—	Dusty
Soil reaction (pH) (surface layer)	—	—	<3.5	Too acid
Sodium adsorption ratio	—	—	>13	Excess sodium
Permeability[c] (cm/hr)(0–1 m)	>1.5	0.15–1.5	<0.15	Slow percolation

[a] Texture abbreviations: LCOS—loamy coarse sand, COS—coarse sand, S—sand, FS—fine sand, LFS—loamy fine sand, VFS—very fine sand, SC—sandy clay, SIC—silty clay, C—clay, SIL—silt, L—loam, VFSL—very fine sandy loam, LS—loamy sand.

[b] Disregard unless soil is in Torr, Arid, or Xer suborder, great group, or subgroup.

[c] Soils in Ust, Tor, Arid, or Xer suborders, great groups, or sub groups rate one class better.

are natural or landscaped tracts used primarily for preparing meals and eating out-of-doors. These areas are subject to heavy foot traffic, and vehicular traffic is confined to access roads and parking lots. The soils are rated on the basis of soil properties that influence development costs of shaping sites or building access roads and parking areas, trafficability, and growth of vegetation after development.

The soil properties of concern would appear to be those providing good trafficability: the surface soil in the picnic area should absorb rainfall readily, remain firm to heavy foot traffic, and not be dusty when dry. The second step is to place quantitative limits on the soil properties that should be considered as restrictive. Table 21.1 illustrates one guide to this soil use. The severity of the limitations is given, as well as a common or general expression of the problem. Perhaps most important, however, are the quantitative critical limits that will document the basis for the interpretation should the soil scientist or user be called upon to defend his or her decisions. In this example, the three categories of limitations—slight, moderate, or severe—are used. The implication is that the relative cost of developing a picnic area would probably be least for locations with slight limitations and greatest for locations with severe limitations. Potential suitability for a specified use generally includes categories such as high, medium, low; or good, fair, and poor.

Alternative representations of the limitations or suitability include a verbal rating or a numerical score. The verbal rating "no limitations" means that the soil has features very suitable for the specified use. A rating of "limitations" means that the soil has some features that are favorable for the specified use and some that are unfavorable. The interpretation identifies the most significant limitations for any given soil. The limitations can be overcome or minimized by special planning, design, or installation. Fair to poor performance and moderate to high maintenance costs can be expected, depending on the number of limitations and the severity of each limitation. When a numerical score is provided, the rating is provided as a decimal fraction between 0 and 1.00. A score of 0 means that the soil feature is not a limitation ("not limited") for the specified use; a score of 1.0 means that the soil feature has the greatest negative impact ("very limited") on the specified use.

Interpretations of Soil Material in Soil Horizons. The following soil properties, qualities, and behaviors are included in most soil survey interpretation tables (USDA-NRCS 2009):

Soil Material Properties: textural class, stone and gravel percent, soil reaction (pH), salinity, high-water table levels, cemented pans, bedrock depths.

Qualities and behaviors: Unified Soil Classification or American Association of State Highway and Transportation Officials (AASHTO) classification of liquid and plastic limits, permeability, available water, shrink–swell, corrosivity, erodibility, subsidence, hydrologic soil group, and potential frost action.

Although soil characteristics important for civil engineering are included in the above list, on-site investigations are necessary on small tracts for which planned use is intense such as construction of heavy buildings, main highways, and large waste disposal sites. This is not only because of the risk of soil spatial variability but also because of the need for licensed engineers to make the interpretations and recommendations for these types of construction. And, as pointed out by Lindsay et al. (1973), soil survey information does not eliminate the need for deep borings supervised by engineers and other specialized tests as part of on-site investigations.

Interpretations of Soil Map Units. In preparing interpretations and giving advice on their use, keep in mind the minimum decision area concept (Doucette 1983). A minimum decision area is the smallest feasible size and shape for which a particular management system or land-use alteration can be applied. For example, it could be that the smallest possible area on which a septic tank system could be installed is about 500 m^2 (0.12 acre), whereas the smallest area for which it is feasible to design a subsurface soil drainage system for agricultural purposes is approximately 12,150 m^2

(3 acres). Interpretations of soil map units consider not only the properties of the named soils in that map unit but also the properties of included soils, the slope of the land, and permanent management technologies such as engineered drainage systems. Interpretations related to soil map unit delineations include crop-forest-range production under specified levels of management, flood frequency, and suitability for recreational facilities and wildlife management.

Examples of Technical Soil Classification Systems. In addition to the soil interpretations described above, each of which can be considered a technical classification system, a number of more comprehensive technical soil classification systems have been developed that focus on specific land uses. We briefly describe three of them here: the USDA Land Capability Classification system, the Storie Index (for agriculture and timber), and the Fertility Capability Classification system.

The Land Capability Classification (LCC) system groups arable lands based on potential and limitations for sustained production of commonly cultivated crops that do not require special site conditioning or treatment (Klingebiel and Montgomery 1966). Nonarable lands are grouped based on their ability to support permanent vegetation and on the risk of soil damage if mismanaged. The classification system incorporates both climate and soils information and rates lands for irrigated and nonirrigated management systems. At the most general level, landscape (map) units are grouped into one of eight classes (I is highest; VIII is lowest). Lands in classes I through IV are arable; lands in classes V through VIII are nonarable.

Class I lands have few limitations for cultivated crop production. Soils are on nearly level parts of the landscape, rainfall or irrigation provides sufficient water, the frost-free season is long enough to allow crops to mature, and salts, flooding, or a high water table do not occur. Class II lands have some limitations, and careful management is required to maintain productivity, but management practices are easy to apply. Class III lands have severe limitations and require special management and crops to maintain productivity. Class IV lands have very severe limitations that generally restrict the range of crops that can be grown and that require intensive management to maintain productivity. Limitations may include, among others, slope, soil thickness, salinity, high water table, and low water-holding capacity. Class V lands have level soils with special management problems (for example, flooding, short growing season, stony and rocky soils, and areas with stagnant surface water that cannot be drained) that limit their use to pasture, range, woodlands, and wildlife habitat. Class VI and VII lands have management problems that restrict use for range, pasture, and woodland. Class VIII lands are nonagricultural lands that are generally used for watershed, recreation, and wildlife.

The next categorical level is the subclass, which groups classes on the basis of similar kinds of limitations. The subclasses include *e* where erosion is the main risk; *w* where wetness caused by impeded drainage or overflow is the main problem; *s* where soil limitations, such as shallow soil, salinity, or low water-holding capacity, are the main problems; and *c* where a climatic limitation is the main problem and when *e*, *w*, and *s* are not appropriate.

The lowest categorical level is the capability unit. Each state within the United States has its own mechanism for identifying the problems that are most limiting. Generally speaking, the soils in one capability unit are enough alike to be suited to the same crops and pasture plant, to require similar management, and to have similar productivity. Soils in Class I do not have a subclass or capability unit because there are no major limitations. The data elements that drive the LCC ratings are found in NASIS. Details of the LCC criteria are found in Exhibit 622-2 in the *National Soil Survey Handbook* (USDA-NRCS 2009).

The Storie Index Rating (Storie 1933) is a quantitative, factorial productivity rating system that was developed for irrigated row crop production in California. This index expresses numerically the relative degree of suitability of a soil for general intensive agricultural uses at the time of the evaluation. The rating is based on soil characteristics and is obtained by evaluating soil surface and subsurface chemical and physical properties, as well as landscape surface features. Not considered in the rating are availability of water for irrigation, local climate, size and accessibility of mapped areas, distance to markets, and other factors that might determine the desirability of growing certain plants in a given location. Therefore, the index cannot be used as the only indicator of land value. Where the local economic and geographic factors are known to the user, however, the Storie index may provide additional objective information for land tract value comparisons.

Four general factors are used in determining the index rating: A—the permeability, available water capacity, and depth of the soil; B—the texture of the surface soil; C—the dominant slope of the soil body; and X—other conditions more readily subject to management or modification by the land user. These other conditions may include drainage, flooding, salinity, alkalinity, fertility, acidity, erosion, and microrelief. For some soils, more than one of these X conditions is used in determining the rating. A rating of 100% expresses the most favorable condition for general crop production. Lower percentage ratings are assigned for less favorable conditions. Factor ratings, in percentages, are selected from tables prepared from data and observations that relate soil properties to plant growth and crop yields. Certain properties are assigned a range of values to allow for variations in the properties that affect the suitability of the soil for general agricultural purposes.

The index rating for a soil component of a map unit is obtained by multiplying the percentage rating values (expressed as a decimal fraction) given to its four factors, A, B, C, and X. If more than one condition is recognized for the X factor for a soil, the value for each condition acts as a multiplier. Thus, any of the A, B, C, or X factor conditions may dominate or control the final rating. If a map unit consists primarily of one named soil series (a consociation), the index rating for the named soil component equals the index rating for the map unit. If a map unit consists of more than one named component (a complex or association), ratings are assigned to each named component (soil series or miscellaneous area, such as “Rock outcrop”), and a weighted map unit index is calculated from the component indexes and the proportion of each of the named components in the map unit. Miscellaneous areas are considered to be unsuited for agriculture, and are assigned a rating of zero. Inclusions of other soils, not named in the map unit name, are ignored in the calculations.

The final decimal product of the mutliplied factors is converted to a percentage value between 0 and 100. Map units are assigned grades according to their suitability for general intensive agriculture as shown by their Storie index ratings. The six grades and their range in index ratings follow:

Grade 1: 80 to 100
Grade 2: 60 to 80
Grade 3: 40 to 60
Grade 4: 20 to 40
Grade 5: 10 to 20
Grade 6: 0 to 10

Grade 1 soils are well suited to intensively grown irrigated crops that are climatically adapted to the region. *Grade 2* soils are good agricultural soils, although they are not as desirable as soils in grade 1 because of a less permeable subsoil, deep cemented layers (such as, duripans), a gravelly or moderately fine textured surface layer, moderate or strong slopes, restricted drainage, low available water capacity, lower soil fertility, or a slight or moderate hazard of flooding. *Grade 3* soils are only fairly well suited to agriculture because of moderate soil depth; moderate to steep slopes, restricted permeability in the subsoil; a clayey, sandy, or gravelly surface layer; somewhat restricted drainage; acidity; low fertility; or a hazard of flooding. *Grade 4* soils are poorly suited. They are more limited in their agricultural potential than the soils in grade 3 because of restrictions, such as a shallower depth; steeper slopes; poorer drainage; a less permeable subsoil; a gravelly, sandy, or clayey surface layer; channeled or hummocky microrelief; a hazard of flooding; or low fertility; salinity or alkalinity, or acidity. *Grade 5* soils are very poorly suited to agriculture and are seldom cultivated. They are more commonly used as pasture, rangeland, or woodland. *Grade 6* soils and miscellaneous areas are not suited to agriculture because of very severe limitations. They are better suited to limited use as rangeland, protective wildlife habitat, woodland, or watershed.

A computerized calculation of the Storie Index has been developed to eliminate some of the subjectivity encountered during manual calculation of the Index (O'Geen and Southard 2005). The calculation is driven by data from NASIS, and utilizes discrete and fuzzy logic functions as a means for assigning numerical scores to soil properties (O'Geen et al. 2008).

Storie and Wieslander (1948) developed a Timber Rating using a similar quantitative, factorial approach focused on commercial conifer species of the Sierra Nevada, Cascades, and northern Coast Ranges of California. Five factors are evaluated: A—soil depth and suitable texture, B—soil permeability, C—soil chemical characteristics (alkalinity, salinity), D—soil drainage and run off, and E—mean annual precipitation. Numerical values are derived from tables based on the relationship between tree growth (height-age site index) and site properties as determined from extensive field measurements. The product is expressed as a percentage (up to 120% to accommodate coastal redwoods). Ratings are assigned based on the index rating percentage:

Redwood: 98 to 120
High: 75 to 98
Medium: 50 to 75
Low: 30 to 50
Nontimber: 0 to 30

Soils rated as *Redwood* are well suited to intensively grown timber that is climatically adapted to the region. *High* soils are good timber soils, although they are not as desirable as soils in the Redwood category because of a less permeable subsoil, deep cemented soils layers, a gravelly or moderately fine-textured surface layer, restricted drainage, low available water capacity, or less rainfall. *Medium* soils are only fairly suited to timber harvesting because of soil depth; restricted permeability in the subsoil; a clayey, sandy, or gravelly surface layer; somewhat restricted drainage; low fertility; flooding hazards; or low rainfall rates. *Low* soils are very poorly suited for timber production and are more commonly used for pasture, rangeland, or recreational purposes. *Nontimber* soils and miscellaneous areas are not suited for timber production because of very severe soil or climatic limitations. They are better suited for use as rangeland, wildlife habitat, or watersheds.

The Fertility Capability Soil Classification (FCC) system is a technical system for grouping soils according to the kinds of problems they present for agronomic management of their chemical and physical properties (Buol et al. 1975; Sanchez et al. 1982). This interpretive system uses readily measurable topsoil characteristics and some subsoil properties important for plant growth. The system is designed to interrelate and tie together the subdisciplines of soil fertility and soil classification by grouping soils according to properties that are related to the response of a soil to fertility management practices. It is best used in conjunction with soil testing practices. The system has three categorical levels: the type (topsoil texture), substrata type (subsoil texture), and 15 modifiers. The modifiers systematically identify soil properties near the soil surface that agronomists use to predict expected responses of individual kinds of soil to fertility related practices. Some examples of modifiers are *a*: aluminum toxicity (such as, >60% exchangeable aluminum saturation of the effective cation exchange capacity; *e*: low (ECEC) within 60 cm of soil surface; *i*: P fixation by iron oxide surfaces; *x*: P fixation by short-range order (amorphous) aluminum-silicates; *g*: reducing conditions and potential denitrification. The system uses quantitative ranges of soil properties but accepts modifications to accommodate localized management technologies. It can be used for regional evaluations of fertility requirements when applied to soil survey map units using *Soil Taxonomy,* FAO world soil maps, or other quantitative systems of soil classification.

Perspective

Soil survey interpretations apply the information and knowledge contained in a soil survey to the evaluation and prediction of the suitabilities, limitations, and potential of soil materials and landscape units for a variety of specific uses. The interpretations are prepared to help land users, public officials, land use planners, scientists, and technologists make better decisions about soil resources. Interpretations are organized in a variety of use-specific technical soil classifications, which rely on a subset of soil properties contained in the natural soil classification system. Soil survey interpretations for the National Cooperative Soil Survey are developed from data contained in the National Soil Information System. This database is populated with actual and derived soil data from pedon descriptions, laboratory characterization, and map unit characteristics. Interpretations are made for a wide variety of soil uses, including agricultural, environmental, and land use-planning purposes. Interpretations are made for both soil materials and areas of land occupied by populations of soils and depicted as map unit delineations on soil maps. The minimum decision area concept is a useful guideline for interpretations of soil map unit delineations. The Land Capability Classification System, the Storie Index Rating, and the Fertility Capability Classification System are examples of technical soil classification systems that use soil and map unit properties to rate soils for agricultural and timber production purposes. Most modern soil survey information is stored in computer databases that are readily accessible online.

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