What Soils Are

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Soil is continuous over the land surface of the earth, except for the steep and rugged mountain peaks and the lands of perpetual ice and snow.

Soil is related to the earth much as the rind is related to an orange. But this rind of the earth is far less uniform than the rind of an orange. It is deep in some places and shallow in others. It may be red, as soils are in Hawaii, or it may be black, as they are in North Dakota. It may be sand, or it may be clay.

Be it deep or shallow, red or black, sand or clay, the soil is the link between the rock core of the earth and the living things on its surface. It is the foothold for the plants we grow. Therein lies the main reason for our interest in soil.

The soil mantle of the earth is far from uniform, but all soils have some things in common.

Every soil consists of mineral and organic matter, water, and air. The proportions vary, but the major components remain the same.

Every soil occupies space. As a small segment of the earth, it extends down into the planet as well as over its surface. It has length, breadth, and depth.

Every soil has a profile—a succession of layers in a vertical section down into loose weathered rock. The nature of the soil profile has a lot to do with the growth of roots, the storage of moisture, and the supplies of plant nutrients. The profile also is basic to scientific studies of soil. The profile carries within itself a record of its history for those who learn to read it.

A SOIL PROFILE consists of two or more layers lying one below the other and parallel to the land surface. The layers are known as horizons. The horizons differ in one or more properties such as color, texture, structure, consistence, porosity, and reaction.

Soil horizons may be thick or thin. Some are no more than a fraction of an inch. Others are several feet thick. Few horizons are at either extreme. Generally they merge with one another and lack sharp boundaries.

Horizons in a profile are like the parts of a layer cake without the clear bands of frosting between them.

Most soil profiles include three master horizons, identified by the letters A, B, and C. Some soils lack a B horizon and are said to have AC profiles. When a soil is used without proper care, the A horizon and even the B horizon may be eroded away.

The combined A and B horizons are called the solum, sometimes the "true soil." Together they form the major part of a profile. They are also direct results of the processes by which soils are formed.

All of the master horizons may be subdivided in the scientific study of soils. Such subdivisions are identified by the proper letter plus a subscript number, thus: A_1, A_2, A_3, B_1, B_2, B_3. The subdivisions of master horizons provide clues to the processes of soil formation and are important to the use and management of soils.

The A horizon, the uppermost layer in the soil profile, often is called the surface soil. It is the part of the soil in which life is most abundant in such forms as plant roots, bacteria, fungi, and small animals. It is therefore the part in which organic matter is most plentiful.

Because it lies at the surface, the A horizon is also the part of the soil that falling rain reaches first. Hence it normally is leached more than are the deeper horizons. Most A horizons have

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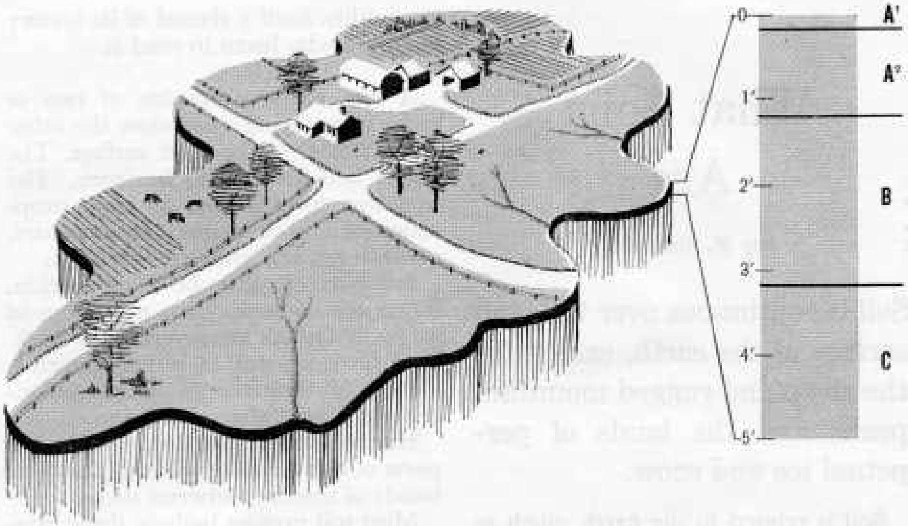


1. A single area of a soil type as it occurs in nature. At the right is an enlarged sketch of the profile with its major horizons.

lost soluble substances. Many have also lost some clay—mineral particles finer than a pinpoint. (About 10 thousand clay particles of average size laid end to end would equal 1 inch.) They may also have lost iron and aluminum oxides, which soil scientists generally call sesquioxides. Iron oxide is the familiar rust on an old piece of steel. Aluminum oxide is the dark tarnish on aluminum kettles.

Two subdivisions of the A horizon are common in soil profiles, although only one of the two may be present in a given profile. If a soil has been formed under prairie vegetation, as in the Corn Belt, it has a thick, dark A_1 and lacks an A_2 horizon. If the soil were formed under forest cover in the same region, the A horizon has two distinct subdivisions, a thin, dark surface layer (the A_1 horizon) and a much thicker, lighter colored layer beneath it (the A_2 horizon). The A_1 and A_2 horizons can be recognized in many uneroded soil profiles in humid regions, but both exist in few soil profiles of dry regions.

The B horizon lies immediately beneath the A horizon and often is called the subsoil. Lying between the A and C horizons, it partakes of the properties of both. Living organisms are fewer than in the A horizon but more abundant than in the C horizon. Color is often transitional between those of the A and C horizons. The B horizon generally is harder when dry (and stickier when wet) than its neighbors. It is frequently higher in clay than either of them. It may have a blocky or prismatic structure, usually combined with greater firmness. Concentrations of iron oxide or aluminum oxides or both, usually in combination with organic matter, mark B horizons of some soils.

THE C HORIZON is the deepest of the three major horizons. It consists of the upper part of the loose and partly decayed rock beneath the A and B horizons. The rock material in the C horizon is of the same kind as that which now forms the bulk of the soil above it. The C horizon therefore is said to be the parent material of soils. It may have accumulated in place by the breakdown of hard rock, or it may have been moved to where it now is by water, wind, or ice. The C horizon has less living matter than overlying ones and is therefore lower in organic matter. It is commonly lighter in color than the A and B horizons. The C horizon in most soils is more like the B than the A horizon. Some profiles, however, lack B horizons. Such profiles usually consist of faint or distinct A horizons grading downward into C horizons. The differences between the A and C horizons may then be small, especially if the A horizon is faint.

The master horizons and their subdivisions, recognized in scientific studies of soils, are shown in the second diagram. This hypothetical profile cannot be found in nature. All the horizons and subhorizons in it do not exist in any actual soil. Yet some of the horizons are part of every soil on earth. Moreover, the kinds and arrangement of horizons in a profile are a record of what has happened to that soil since it began to form. This history has meaning to the fertility, tilth, and productivity of soils for plants useful to mankind.

Soil formation proceeds in steps and stages, none of which is distinct. They are like the overlapping fibers in a piece of string—the eye can hardly tell where one fiber ends and another begins. Similarly, it is not possible to be sure where one step or stage in soil formation stops and another starts.

The two major steps in the formation of soils are accumulation of soil parent materials and differentiation of horizons in the profile. Each step can be thought of as consisting of several stages, which are hard to tell apart.

Some processes begin with the onset of rock weathering. Silicate clay minerals, for example, are formed and oxides are released from primary minerals, such as feldspars, before a rock disintegrates. Feldspars are minerals formed when molten lava crystallizes into hard rock, as happens deep in the earth. They consist of aluminum, silicon, and oxygen and one or more of calcium, potassium, and sodium. Silicate clay minerals are also mostly silicon, aluminum, oxygen, and hydrogen, but differ in atomic structure from feldspars.

Clay minerals may continue to be formed and oxides released within a soil as long as it exists. Minerals normally continue to decompose in a soil profile long after the distinct horizons form. Some processes in horizon differentiation, however, may begin only after there is a deep mantle of loose weathered rock. The two main steps in soil formation thus merge with one another.

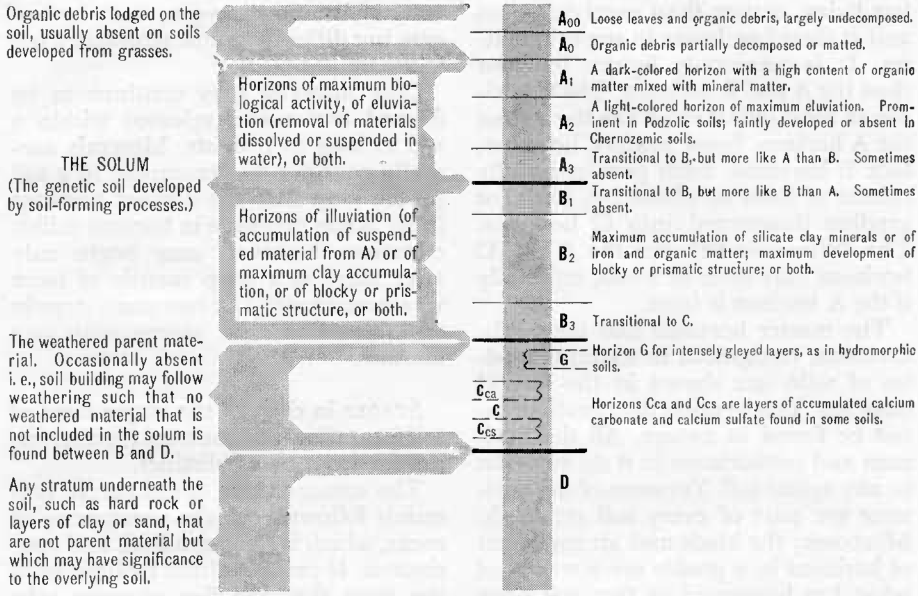
Stages in each of the major steps of soil formation, like the steps themselves, are far from being distinct.

The accumulation of soil parent materials follows from the weathering of rocks, which is slow, gradual, and continuous. It proceeds little by little from the time that the first changes take place in the solid rock mass. Many changes normally occur in a rock before it disintegrates. Decomposition of minerals usually proceeds long after a rock has disintegrated. Weathering continues without any sharp breaks. Thinking of it in stages is simply a way of looking at a continuous process one piece at a time.

Horizons in soil profiles may be considered faint, distinct, and prominent, although the stages in their differentiation are not clearly defined. Two or more faint horizons first appear when a soil profile begins to form. The earliest changes are small, and the horizons therefore are faint and hard to distinguish. If conditions are favorable, these horizons become distinct with the passing of time.

Additional horizons may also appear. Horizons change slowly and gradually as they become more and more distinct. These changes are like the steady upward movement of a barrel being rolled up an incline rather than like the jumps with which a boy goes upstairs.

Changes in color, structure, texture, consistence, porosity, and other properties occur in all horizons, but the



2. *A hypothetical soil profile that has all the principal horizons. Not all of these horizons are present in any profile, but every profile has some of them.*

same changes do not take place in every one. These various kinds of changes slowly differentiate horizons in a soil profile.

THE WEATHERING OF ROCKS provides soil parent materials.

Solid rocks disintegrate slowly under the influence of climate, which acts on them through sunshine, rain, frost, and wind. Heating and cooling, freezing and thawing, wetting and drying all tend to weaken the rock structure. The minerals in rocks react with water and air that enter through tiny cracks and crevices. Changes in the minerals then set up stresses and strains, which further weaken the rock structure.

The final effect of these forces is to break up a rock into small pieces, often into the constituent mineral grains. Gradually rocks thus disintegrate and decay. The loose and weathered rock materials may then become soil parent materials. As used here, parent rock means rock that is still solid and massive, whereas soil parent material is the distintegrated rock at or near the present land surface.

Tremendous quantities of rocks have been weathered during the millions of years that have passed since the continents took form. A mantle of weathered rock, known as the regolith, now blankets the land surface generally. This regolith has been formed in some places by disintegration and decomposition of rocks on the spot. In many more places it has been moved about by water, wind, or ice.

The regolith may be slightly weathered and consist of fresh primary minerals, as it does in recent deposits left by glaciers in Alaska. It may be intensely weathered so that nothing but highly resistant minerals remain, as is true in parts of central Brazil. The wide range in degree of weathering is matched by similar ranges in composition of the regolith and in its thickness to hard rock. The composition and thicknesses are important to soil formation. The nature of the original rock and the stage of weathering of the regolith also affect the fertility and water relationships of soils.

PLANTS SOON GAIN A FOOTHOLD in the regolith. Sometimes they begin growing on rock before it has disintegrated. The pioneers are commonly simple forms, such as lichens. Micro-organisms, such as bacteria and fungi, also are early invaders. Larger and more complex plants soon follow. Small animals then join the biological community in the infant soil.

As these organisms grow and die, their bodies are left on and in the regolith. Parts of dead plants fall to the surface. Roots are left within the weathering rock mass. The addition and decay of organic matter gradually change the character and appearance of the surface layer of the regolith. It begins to differ from the deeper layers. It thus becomes a faint A horizon, marking the first stage in the differentiation of horizons.

Formation of an A horizon usually follows on the heels of weathering. By the time a regolith has been formed, horizon differentiation also will have started. Soil profiles with A and C horizons, both of which may be thin, therefore can be found in all but the very youngest regoliths. With the passing of time, a faint A horizon slowly becomes thicker and plainer, growing at the expense of the C horizon.

The B horizon makes its appearance after the A horizon has become distinct, as a rule, although A and B horizons may be formed together. Some soils have AC profiles for a long time before there is any indication of a B horizon.

Many soils in north central Iowa have thick, dark A_1 horizons with little evidence of any B horizons. These soils were formed from drift left by glaciers some 8,000 years ago. Eighty centuries have passed since horizon differentiation began, and an A_1 horizon probably has been obvious most of that time.

A_2 and B horizons have been formed in the former plowed layers of fields abandoned in eastern North Carolina a scant half century ago. These fields are now covered by pines about 45 years old. Under the trees, leaching has formed a thin A_2 horizon, and the accumulation of iron oxides and organic matter has formed a thin B horizon. The combined thickness of the two horizons is no more than 5 inches. Both of these new horizons lie within what was the A_2 horizon of a larger profile formed earlier. Changes reflected in the new profile are small, even though the horizons are distinct.

HORIZONS ARE FORMED in soil profiles because of gains, losses, and alterations.

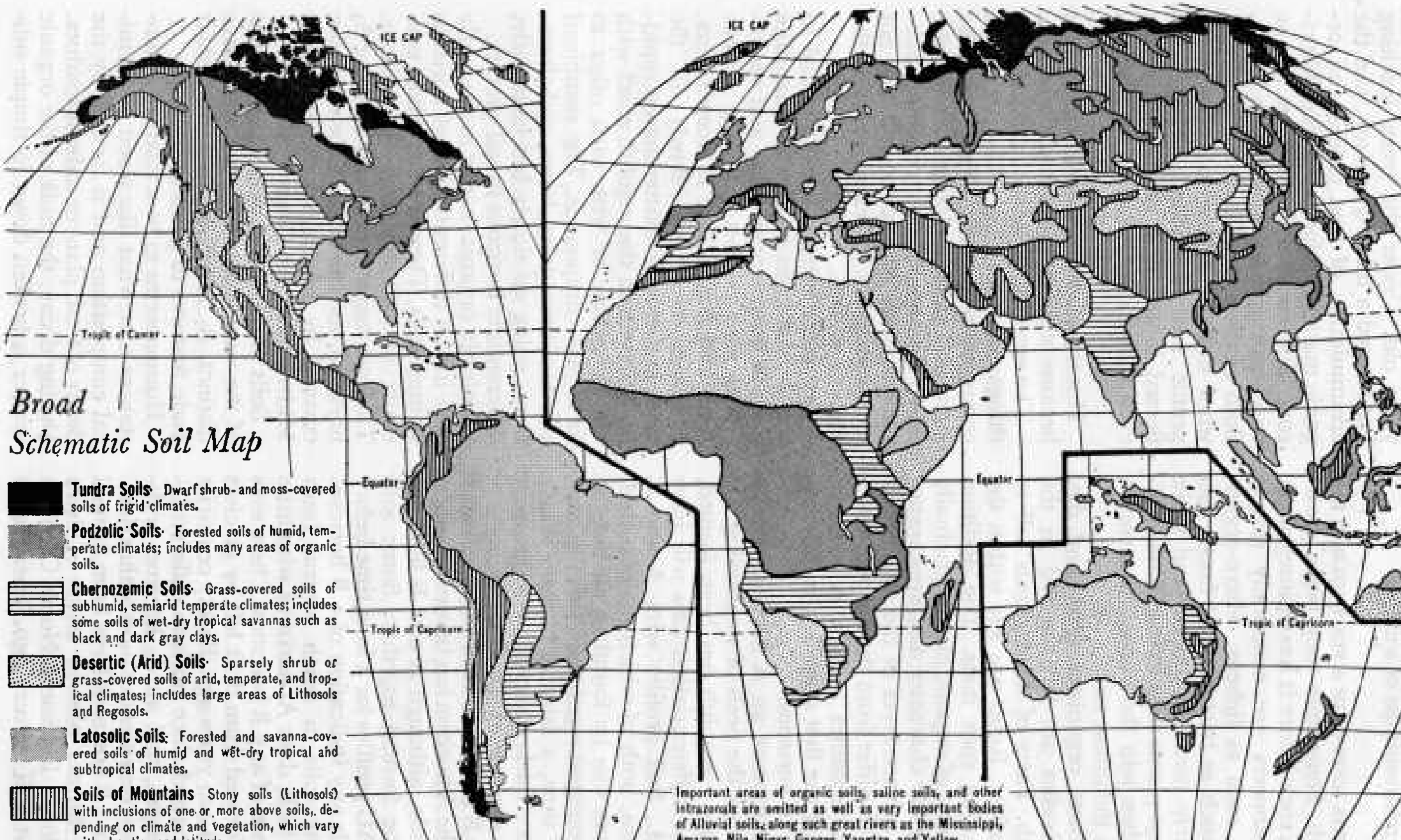
The regolith, as a whole, gains some things and loses others. Changes in its composition go on all the time. The kinds and rates of changes are not the same throughout the regolith. Substances being gained or lost are not identical in the different parts of the vertical section.

Organic matter is usually added to the surface layer in greater quantities than to deeper ones. Clay and sesquioxides may be lost from the surface layer and accumulate in deeper ones. Minerals decompose slowly all the time, and organic matter decays rapidly. Minerals in the regolith react with air and water or with each other to form new compounds. The combined effects of these gains, losses, and alterations, going on slowly but constantly, differentiate horizons in soil profiles.

Gains in organic matter are an early step in the differentiation of horizons in most soils. They are important generally in differentiating A horizons. They are major processes in forming A_1 horizons, such as those in the soils once covered by prairie vegetation in the Midwest and the Great Plains.

Soils of the broad chernozemic belts of the world, shown in the map, have distinct A_1 horizons that owe their main features to gains in humus.

Humus seems to be a mixture of substances somewhat like wood in chemical composition and is formed by the decay of fresh plant or animal residues. During decay, the bulk of the organic matter is broken down to simple sub-



3. Map of the world, showing six broad soil zones. Each zone generally has similar processes of horizon differentiation prevailing over it. These are reflected in the character of the well-drained soils with undulating to rolling topography. Many kinds of soils are present in every zone.

stances, such as water and carbon dioxide. (Gas bubbles in soft drinks are carbon dioxide.) Black colors, as in chernozemic soils, do not always follow from additions of organic matter. Moderate amounts are added throughout the profile to many soils in the latosolic belts, mainly in the Tropics, but the dominant colors are still red and yellow.

Being the remains of living organisms, soil organic matter has nutrient elements in its structure and thus becomes a small storehouse for such elements. It also has direct effects on structure and tilth.

Gains in organic matter are not generally important in forming B horizons. Small amounts are transferred downward from A to B horizons in many soils. Usually the effects in horizon differentiation also are small. Large amounts of organic matter, as humus and more soluble forms, are moved down from the A horizon of some podzolic soils formed from sands. Thus much organic matter is being lost by the A and gained by the B horizon in such soils. The humus moves largely in suspension, tending to stop in the B horizon. More soluble organic matter moves in solution, and part of it may also stop in the B horizon.

Losses from the soil profile are due mainly to leaching or eluviation. Downward movement of substances in solution is known as leaching. The washing out of mineral and organic matter in suspension is called eluviation. The reverse of eluviation is illuviation, which means "washing in" rather than "washing out." The reverse of leaching is deposition. Losses may also be due to removals of nutrient elements by plants.

Water must move through a soil profile before leaching and eluviation can occur. The water comes from falling rain, which first enters the A horizon. In passing through the atmosphere, the water has absorbed a little carbon dioxide, which makes it slightly acid and helps it to dissolve minerals.

Once in the soil, the water dissolves minute quantities of mineral and organic matter, as sugar dissolves in coffee. The dissolved substances commonly move with the water as far as it goes. To this general rule, there are exceptions.

Material in solution may be carried deeper in the profile, below that into the regolith, or out into streams and eventually to the sea. If enough moves through the soil, water may also pick up clay, humus, or sesquioxides, as in muddy streams. Losses through leaching and eluviation and gains through illuviation are important to horizon differentiation in many soils.

Leaching and eluviation affect soils of humid regions more, on the whole, than they do those of dry regions. The rainfall is greater, so more water moves down and less moves upward through the profile. Examples of leached and eluviated A_2 horizons are common in soils formed under forest in the Eastern States and in the broad podzolic belts of the world.

Soluble salts (sodium chloride, or table salt, is an example), carbonates, and elements such as calcium, sodium, and potassium are slowly leached from the profiles of well-drained soils in humid regions. Leaching gradually removes the more soluble elements and also carries out some part of the less soluble ones, such as silicon, from most soils of the podzolic and latosolic belts. Slowly but surely, leaching depletes the plant nutrients in these soils. In extreme cases, among soils of the Tropics, the full supply of available nutrients is bound up in living and dead organic matter.

Leaching also plays a role in horizon differentiation in soils of dry regions. Soluble salts and carbonates in the well-drained soils are leached from the A horizon and also may be removed from the B horizon. Carbonates may be left in the lower B horizon but are left oftener in the C horizon.

Soluble salts, as a rule, are moved down and out of the profiles of well-drained soils. They have been leached to depths of 8 to 10 feet in the glacial till under many soils along the Canadian border in North Dakota. The A and B horizons of soil profiles in that region have combined thicknesses of 2.5 feet. In soils without free drainage, on the other hand, soluble salts and carbonates accumulate wherever the water table may stand in a profile, even at the surface itself. Generally speaking, leaching of A horizons is greater than that of B horizons in both arid and humid regions.

Eluviation operates mainly on A horizons. Clay, humus, and sesquioxides are lost from the A horizons of many soils. A part of these substances is carried out of the profile into the drainage waters. Mostly, however, the losses from the A horizon by eluviation are offset by gains in the B horizon. Many soils in humid regions show the effects of eluviation of substances from the A horizon accompanied by illuviation in the B horizon. Some soils in dry regions also bear the marks of eluviation and illuviation in their profiles.

Changes in the composition of mineral and organic matter are normal in soils the world over. Some of the changes proceed slowly. The decomposition of minerals is slow. The decay of organic matter is rapid. Whether fast or slow, alterations in composition continue throughout the regolith, including the soil profile. Changes that take place in a single week may be negligible, but Nature is patient and has plenty of time. The decomposition of minerals can become profound after many thousands of years.

Soil organic matter turns over rapidly. Most plant residues added to the A horizon by the growth of one season will decay before the next has come and gone. Because fresh supplies are added periodically, organic matter in the soil is in all stages of decay. A very small part of the total decomposes slowly enough to persist for several centuries. Some products of decay, such as humus, have direct effects on color, structure, and consistence of soil horizons. Others react with minerals to speed up their decomposition and change. Some will dissolve sesquioxides, which can then move more rapidly. The decay of organic matter also releases nutrient elements that may later be used for plant growth.

Alterations of minerals and organic matter may make them more soluble or less soluble, or the solubility may be unchanged. Elements such as calcium, sodium, and potassium are shifted to more soluble forms when feldspars decompose. They then may be absorbed by roots or lost through leaching. Iron is released by decomposition of minerals much like (but not identical with) feldspars. It often combines with oxygen soon afterward to form oxides or rust, which is almost insoluble.

The breakdown of feldspars with formation of clay minerals has little effect on solubility. Both types of minerals are slightly soluble. The individual clay particles are smaller than feldspars, however, and move in the soil more readily.

Changes in the composition of minerals and organic matter may contribute directly or indirectly to horizon differentiation. When clay and sesquioxides are formed by decomposition of feldspars and other primary minerals, this in itself can make one horizon differ from another.

Differences in amounts of clay between A and B horizons in many soils reflect in part the differences in rates of formation of clay minerals in each of the two horizons. Changes in composition may also shift the solubilities of substances. This can then speed up or slow down leaching, eluviation, illuviation, and deposition.

Gains and losses from the regolith, especially from the soil profile, coupled with alterations in minerals and organic matter, determine the direction and rate of horizon differentiation.

They, in turn, depend on a host of simple processes, such as hydration, hydrolysis, solution, leaching, eluviation, and illuviation. These simple processes prevail in all soils. The balance is governed by such factors as climate, vegetation, and parent rock. The total combined effects of the simple processes fix the character of every soil, including its profile and its suitability for the growth of plants.

THE FIVE MAJOR FACTORS in soil formation are climate, living organisms, parent rocks, topography, and time. They control the weathering of rocks and the gains, losses, and alterations throughout the regolith, including the soil profile.

Temperature and rainfall govern rates of weathering of rocks and the decomposition of minerals. They also influence leaching, eluviation, and illuviation. Thus climate functions directly in the accumulation of soil parent materials and in differentiation of horizons. The indirect effects of climate are through its controls over the kinds of plants and animals that can thrive in a region. These living organisms, in turn, are of major importance in differentiating horizons in soils.

Because climate is so important to soil formation, the broad soil regions of the world tend to follow the distribution of climates. Soil and climatic regions are not identical, however, because five factors rather than one are important in soil formation.

Living organisms—plants, animals, insects, bacteria, fungi, and the like—are important chiefly to horizon differentiation and less so to the accumulation of soil parent materials. Gains in organic matter and nitrogen in the soil, gains or losses in plant nutrients, and changes in structure and porosity are among the shifts due to living organisms. Plants and animals may also mix horizons and thus retard their differentiation.

Plants largely determine the kinds and amounts of organic matter that go into a soil under natural conditions. They also govern the way in which it will be added, whether as leaves and twigs on the surface or as fibrous roots within the profile.

Some plants take their nitrogen from the air and add it to the soil as they die. Deep-rooted plants reverse leaching processes in part. The roots may take up calcium, potassium, phosphorus, and other nutrient elements from the C horizon or even from the deeper regolith, only to leave some part of those nutrients in the solum when the plants die.

The effects of plants on the soil beneath them may be striking. Desert shrubs, such as shadscale, for example, concentrate sodium in the soil on which they grow. Enough sodium is taken up by the plant and added to the surface of the soil to make it far more alkaline than the soil between the bushes. Most effects are less dramatic than this, but all are important to horizon differentiation.

Horizons may be mixed by plants or animals. When trees tip over in a forest, the roots take up soil materials from several horizons. As the upturned roots decay, this soil material tumbles back down, mixing as it goes. Burrowing animals also mix horizons as they build their homes. Such mixing partly offsets horizon differentiation.

Bacteria and fungi live mainly on plant and animal residues. They break down complex compounds into simpler forms, as in the decay of organic matter. It has been suggested that the humus in soils is largely dead bodies of micro-organisms; much of it seems to have about the same composition, even though it exists under widely different types of vegetation. Some micro-organisms fix nitrogen from the atmosphere and thus add it to the regolith in their bodies when they die.

Parent rock is sometimes called a passive factor in soil formation. It must be weathered to form soil parent materials, which are further changed as horizons develop in a soil profile.

The character of the rock itself is a factor in the kinds of changes and how fast they go. Pure quartzite will disintegrate, for example, but little else can happen to it. Quartzite consists of quartz grains cemented together by silica. Quartz is also silica, a combination of silicon and oxygen. Highly resistant to weathering, quartz grains are well-nigh permanent. Small ones may dissolve very slowly, but no plant nutrients are released and no clay is formed as they do.

Most rocks are mixtures of many minerals, few of which are able to withstand weathering as well as quartz. The composition and structure of rocks strongly influence the rate of weathering and the products of that weathering. These in turn are both important to the kind of soil that may be formed.

Topography, or the lay of the land, affects runoff and drainage. Other things being equal, runoff is large on steep slopes and small on level ones. Drainage is rapid from mountainsides and slow from level plains. The amount of water that moves through the soil depends partly on topography. More water runs off and less enters the soil on steep slopes, as a rule, than it does on gentle slopes. The runoff also removes more of the weathered rock on steeper slopes, other things being the same.

Soil profiles on steep slopes generally have indistinct horizons and are shallower than those on gentle slopes. Low and flat topography often means that extra water is added to the soil. The extra water is reflected in gray or mottled colors or in higher amounts of organic matter in the A horizon. If water stands on the surface, peat deposits may be formed. Topography thus influences the moisture regime in soil and the erosion from its surface.

Time is required for soil formation—how much time depends on where the processes must start.

A tremendous interval is needed for development of soil from freshly exposed and fairly pure limestone. The limestone dissolves slowly while the rains come and go. As the mass of the limestone dissolves and is carried away, any impurities originally present are left to form a regolith. Millions of years may pass before parent materials have accumulated and horizons have been formed under such circumstances.

Much more time is needed, generally speaking, for the accumulation of soil parent materials than for the differentiation of horizons in the profile. This would be of the first importance if soil formation had to start from scratch. Because of weathering during past geologic time, however, a regolith now exists widely over the continents. In fact, soils have been formed on most land surfaces, perhaps many times, since molten lava first crystallized into rock about 2 billion years ago.

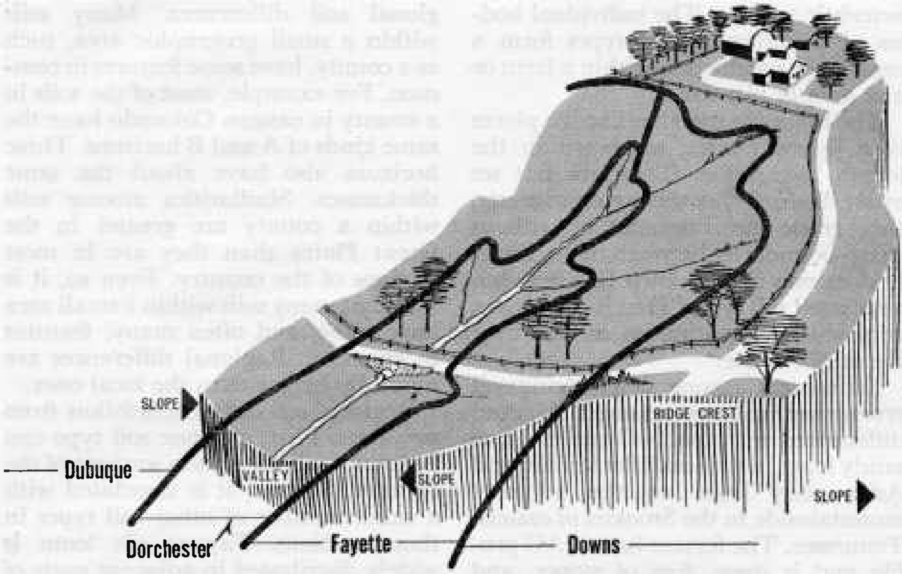
A soil profile may be formed in a fresh regolith within a few decades. R. L. Crocker and J. Major, soil scientists at the University of California, found the topmost 6-inch layer darker and higher in organic matter than the drift below within 30 years after it was left by a melting glacier in southeastern Alaska. Biological activity is low because of the cold climate, yet an A horizon was evident in 30 years. Where the drift has been exposed for 100 years, the A horizon contains as much organic matter as do many in soils of the eastern United States. The leaching of carbonates and increased acidity of the A horizon could also be measured by the end of the first century.

Differentiation of thin A_2 and B horizons within 50 years has also been observed in former plow layers in eastern North Carolina.

Soil profiles usually must be many centuries old before they have distinct B horizons.

Soils with B horizons high in humus have been formed in sands laid down on Roman ruins in parts of western Europe. The sands can be no more than 2,000 years old, but the profiles have prominent B horizons. Conditions for horizon differentiation in the sands were more favorable than they are generally. Furthermore, B horizons high in humus form more rapidly than do B horizons high in clay.

Soils formed from loess—windblown silty sediments—in eastern Iowa have distinct B horizons formed in about 20,000 years. Clay has accumulated in the B horizons of these soils, as have small amounts of sesquioxides.



4. A sketch showing how bodies of soil types fit together in a small landscape, much like the pieces in a jigsaw puzzle. Boundaries between adjacent bodies are gradations rather than sharp lines.

Many soils with B horizons that are thick and either distinct or prominent must be much older than 200 centuries. Long intervals may therefore pass while horizons are being formed in soil profiles, even though parent materials are at hand for horizon differentiation.

None of the five factors of soil formation is uniform over the face of the earth. Variations in all of them are wide.

There are really many climates, many combinations of living organisms, many kinds of rocks, many topographies, and many different ages of land surfaces. As a result, there are hundreds of thousands of different local combinations of the factors of soil formation.

Differences among soils are local and regional.

Every farm consists of several local kinds of soils, known as soil types. A single farm usually has three to six types of soil within its boundaries. Soil types in the whole country number in the tens of thousands. Over the world, they must number in the hundreds of thousands.

These thousands of soil types are not scattered about without rhyme or reason. Each one occurs in a definite geographic area and in certain patterns with others. Thus there are regional differences of importance.

Individual areas of soil types, commonly called phases, are three-dimensional bodies. They have length, breadth, and depth. A single body of a soil type is seldom large. Most are a few acres in size. Each soil type occurs on the earth as a number of separate bodies, which may be distributed over parts of several States or restricted to a small part of one. Each soil type can be defined by a description of the typical profile, the allowable deviations from that profile, and other features such as slope, stoniness, and physiographic position.

Every soil type has neighbors. It never occurs by itself. It tends to have the same neighbors wherever it may be found. A few other soil types usually are associated with it to form a characteristic pattern. The individual bodies of neighboring soil types form a mosaic, or patchwork, within a farm or any area of like size.

The bodies fit together like the pieces of a jigsaw puzzle, as shown in the fourth illustration. They are not set apart distinctly, as the sketch suggests; they grade into one another without sharp boundaries between them. Rarely does one find a sharp line between associated soil types. That is one reason why soil is a continuous mantle over the land surface.

Differences among neighboring soil types may be large or small. Marked differences set apart Congaree fine sandy loam in the small flood plain and Ashe stony loam on the adjacent mountainside in the Smokies of eastern Tennessee. The former has an AC profile and is deep, free of stones, and nearly level. It is fertile and has excellent moisture relationships. Ashe stony loam has faint A, B, and C horizons and is steep, stony, and of low fertility. It has poor moisture relationships because so much of the profile is made up of stones. Neighboring soil types thus may differ greatly in features that have an important bearing on use and management.

Differences among soil types in the same landscape may be small. Fayette silt loam and Downs silt loam occur on the same ridge in northeastern Iowa.

The pattern formed by Fayette, Downs, and associated soils is shown in the fourth drawing.

All soils shown in it might be found on a single farm, but every farm does not include all of them. Fayette silt loam and Downs silt loam have the same horizons in their profiles and have been formed from the same loess. Downs silt loam has thicker A_1, thinner A_2, and slightly more friable B horizons than Fayette silt loam. Water relationships are about the same in the two soils. Downs silt loam is higher in organic matter and a trifle more fertile.

Differences among associated soil types generally are smaller than regional soil differences. Many soils within a small geographic area, such as a county, have some features in common. For example, most of the soils in a county in eastern Colorado have the same kinds of A and B horizons. These horizons also have about the same thicknesses. Similarities among soils within a county are greater in the Great Plains than they are in most sections of the country. Even so, it is true that many soils within a small area have some, and often many, features in common. Regional differences are normally bigger than the local ones.

Regional soil differences follow from two facts. First, any one soil type can be found only in certain sections of the country. Second, it is associated with a small number of other soil types in those sections. Fayette silt loam is widely distributed in adjacent parts of Illinois, Iowa, Minnesota, and Wisconsin. It is not to be found in other States. Congaree fine sandy loam is scattered through the Appalachians and nearby Piedmont from Alabama to Pennsylvania. The two soil types never occur together, and each has its own set of neighbors.

Regional soil differences can be related oftenest to the distribution of climate and living organisms. In places, however, such differences reflect differences in topography, ages of land surfaces, or character of parent rocks. The kinds of regional differences and their size can best be illustrated by comparisons of a few broad classes of soils, each made up of several great soil groups.

A GREAT SOIL GROUP consists of many soil types whose profiles have major features in common. Every soil type in any of them has the same number and kinds of definitive horizons, although they need not be expressed in every profile to the same degree. The Fayette, Dubuque, Downs, and Quandahl soils, for example, are all members of a single great soil group.

The thousands of soil types in the United States can be classified into about 40 great soil groups. For the world as a whole, the number is perhaps half again as large. Collectively, the groups have wide ranges in their many characteristics, or properties.

They also have wide ranges in such qualities as fertility, tilth, ability to hold available moisture, and susceptibility to erosion.

Well-drained soil types with undulating to rolling topography tend to be similar over broad geographic belts. The similarities may extend across from one great soil group to another. Geographic belts marked by certain combinations of great soil groups can therefore be shown on maps of small scale.

Six broad belts are outlined on our schematic soil map of the world.

One consists of mountains and similar rough landscapes, in which many of the soils are stony, shallow, or both. The local patterns of great soil groups and soil types are especially complex in such areas.

The other five broad belts have simpler patterns, but each includes a number of great soil groups. For that matter, the soil types within a single farm commonly represent two or more great soil groups. The broad regions in the map therefore show major kinds of soils rather than complete patterns. Great soil groups other than the dominant ones exist in every belt. These are also important in many localities. The broad belts themselves have much meaning to farming and forestry.

The tundra region has a cold climate, which restricts biological activity and horizon differentiation. The soils are in cold storage for a large part of each year. The deep regolith is permanently frozen in some parts of the tundra. Well-drained soils of the tundra belt have profiles much like those of podzolic soils, although the horizons usually are thinner and less distinct.

PODZOLIC SOILS dominate a broad belt in the higher latitudes of the northern hemisphere and some smaller areas in the southern half of the world. They include the great soil groups known as Podzols (originally from Russian words meaning "ash beneath," referring to the A_2 horizon), Brown Podzolic soils, Gray-Brown Podzolic soils, and Gray Wooded soils. These groups were formed under forest vegetation in humid, temperate climates.

Podzolic soils commonly have distinct A_2 horizons. Some have B horizons that are accumulations of sesquioxides, humus, or both. Others have B horizons that are mainly accumulations of clay with minor amounts of sesquioxides and humus. Podzolic soils are more strongly weathered and leached than chernozemic or desertic soils but less so than latosolic soils. They are commonly acid, low in bases such as calcium, and low in organic matter. Levels of fertility therefore are moderate to low. Available moisture capacity is variable, depending on depth of soil and textures of horizons. As a group, however, the soils are responsive to scientific management.

LATOSOLIC SOILS dominate equatorial belts of Africa and South America. They are also dominant in southeastern parts of Asia and North America, as well as northeastern Australia and the larger islands of the western Pacific Ocean.

They include the great soil groups known as Laterites, Reddish-Brown Lateritic soils, Yellowish-Brown Lateritic soils, Red-Yellow Podzolic soils, and the several kinds of Latosols.

Red-Yellow Podzolic soils are so named because they have some features in common with each of the broad latosolic and podzolic groups, though they are more closely related to the former. Latosolic soils have been formed under forest and savanna vegetation in tropical and subtropical and humid to fairly dry climates. They do not extend into arid regions but may be found in alternately wet and dry zones where rainfall is low.

Latosolic soils are strongly weathered and leached, usually to great depths.

In fact, they are the most strongly weathered soils in the world. Despite the deep, strongly weathered regoliths, most of the soils lack distinct horizons, except for a darkened surface layer or A_1 horizon. Below this, the profile may remain unchanged for many feet. Red-Yellow Podzolic soils differ from others in having distinct A_2 horizons as well.

Red and yellow profile colors are common to latosolic soils because of the large amounts of iron oxides formed through intense weathering. Supplies of plant nutrients normally are low, but the capacity to fix phosphorus in unavailable forms is high. Most latosolic soils are easily penetrated by water and plant roots and are resistant to erosion. Red-Yellow Podzolic soils are more susceptible to erosion than others, being less permeable on the whole. Ease and depth of penetration by roots is illustrated by reports of tree roots extending down to 60 feet in the regoliths beneath latosolic soils in southeastern Brazil.

High porosities are also evident in the many fine tubular channels in such soils. Available moisture capacities are mostly moderate to high in latosolic soils, although they are low in some. Productivity is normally low when latosolic soils are used without benefit of modern science and industry.

CHERNOZEMIC SOILS have been formed under prairie or grass vegetation in humid to semiarid and temperate to tropical climates. These soils are most extensive in temperate zones, but some areas in the Tropics are also large. They include the great soil groups known as Chernozems, Brunizems or Prairie soils, Reddish Prairie soils, Chestnut soils, and Reddish Chestnut soils in temperate regions. In tropical and subtropical regions, the soils have been known as black cotton soils, Grumusols, regurs, and dark clays.

Chernozemic soils normally have dark A_1 horizons of great thickness, are fertile, and but slightly weathered. The A_1 horizons are among the most prominent found in soils, whereas the B horizons usually are much less distinct. The A_1 horizons of chernozemic soils are typically high in organic matter and nitrogen in temperate zones but not in tropical and subtropical zones. The profiles compare in depth to those of podzolic soils. They are not so deep as those of latosolic soils but are deeper than those of desertic soils. Chernozemic soils are less acid, higher in bases, and higher in plant nutrients generally than are podzolic or latosolic soils. They are much higher in organic matter and nitrogen, less alkaline, and lower in bases than desertic soils. Available moisture capacities of the soils are usually moderate to high.

Chernozemic soils of temperate zones are among the naturally most fertile soils in the world. They produce about 90 percent of the grain in commercial trade channels. Within the United States, they form the heart of the Corn Belt and wheat-producing regions. Production varies with seasonal weather, because the soils extend from the margins of humid into semiarid zones.

Chernozemic soils of tropical and subtropical zones commonly have unfavorable physical properties for tillage and plant growth. They are high in clay, plastic, and subject to great shrinking and swelling. Most of them are used for agriculture without benefit of modern technology, and their productivity under simple management is low. Problems in handling the soils are difficult.

DESERTIC SOILS have been formed under mixed shrub and grass vegetation or under shrubs in arid climates. The climates range from hot to cold. The soils are prominent in the great deserts of Africa, Asia, and Australia and in the smaller ones of North America and South America.

They include the great soil groups known as Desert soils, Red Desert soils, Sierozems, Brown soils, and Reddish-Brown soils.

Besides the desertic soils themselves, the deserts of the world include large proportions of sandy wastes, rocklands, and very shallow soils with the barest beginnings of horizons. Such lands lack agricultural possibilities.

Desertic soils have been very slightly weathered and leached. The shortage of moisture which restricts weathering and leaching also limits plant growth, leaving the soils low in organic matter and nitrogen. Limited rainfall is also reflected in the shallow profiles normal to the soils.

Their horizons are seldom prominent. Most of them are faint. The A horizon has commonly lost carbonates and perhaps some bases and clay and is lighter in color than the B horizon. The slightly darker B horizon has some accumulation of clay, but it is very low in organic matter. Levels of nutrient elements other than nitrogen are usually moderate to high in the soils. Available moisture capacities are variable, depending on thickness of profile and textures of horizons.

Marked contrast exists in the productivity of desertic soils used in a highly developed agriculture or under nomadic grazing—an example of the tremendous impact management may have on productivity.

Beyond the effects of management, however, each soil type or great soil group has an ill-defined range in use and management possibilities under a given agricultural technology. Each also differs in its response to changes in technology. The range in possibilities may be narrow, regardless of technology, as it is for steep slopes of Ashe stony loam in the Smokies. Soils of that kind will produce some forest but are not suitable for pasture or crops. On the other hand, Congaree fine sandy loam has a wide range in use and management possibilities. It is suitable for forest, pasture, and a variety of crops within the present agriculture. It would also be productive under a number of other levels of agricultural technology. Yields obtained from Congaree fine sandy loam in any agriculture depend greatly upon the level of management practiced.

Physical Properties

M. B. Russell

The physical properties of a soil largely determine the ways in which it can be used. On the size, shape, arrangement, and mineral composition of its particles and the volume and form of its pores depend other important physical properties.

The flow and storage of water, the movement of air, and the ability of the soil to supply nutrients to plants are examples of properties determined by the size and arrangement of the soil particles.

The proportions of the four major components of soils—inorganic particles, organic material, water, and air—vary greatly from place to place and with depth. The amount of water and air in a soil often fluctuates widely from season to season. The physical characteristics of the primary solid components of soil, however, are essentially unchanging.

Inorganic soil particles occupy about one-half of the total volume of most surface soils. Some of the particles can be seen very easily, but others can be examined only with an electron microscope. For many purposes it is convenient to divide the particles into size groups called separates.

Particles more than 2.0 millimeters in diameter are classed as gravel or stones and are not usually included in analyses of particle size. Particles under 2.0 mm. are divided into three major separates, each of which may be further subdivided.