

Soil

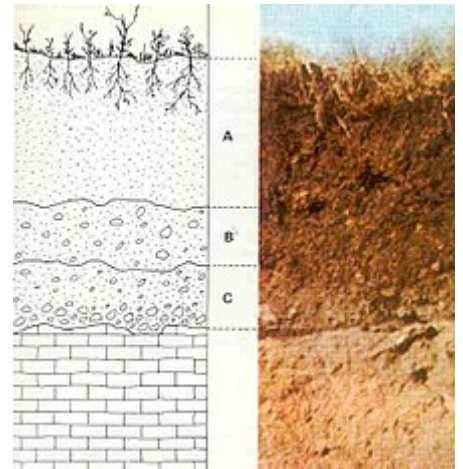
Soil is a mixture of organic matter, minerals, gases, liquids, and organisms that together support life. The Earth's body of soil is the pedosphere, which has four important functions: it is a medium for plant growth; it is a means of water storage, supply and purification; it is a modifier of Earth's atmosphere; it is a habitat for organisms; all of which, in turn, modify the soil.

Soil interfaces with the lithosphere, the hydrosphere, the atmosphere, and the biosphere.^[1] The term *pedolith*, used commonly to refer to the soil, literally translates *ground stone*. Soil consists of a solid phase of minerals and organic matter (the soil matrix), as well as a porous phase that holds gases (the soil atmosphere) and water (the soil solution).^{[2][3][4]} Accordingly, soils are often treated as a three-state system of solids, liquids, and gases.^[5]

Soil is a product of the influence of climate, relief (elevation, orientation, and slope of terrain), organisms, and its parent materials (original minerals) interacting over time.^[6] It continually undergoes development by way of numerous physical, chemical and biological processes, which include weathering with associated erosion. Given its complexity and strong internal connectedness, it is considered an ecosystem by soil ecologists.^[7]

Most soils have a dry bulk density (density of soil taking into account voids when dry) between 1.1 and 1.6 g/cm³, while the soil particle density is much higher, in the range of 2.6 to 2.7 g/cm³.^[8] Little of the soil of planet Earth is older than the Pleistocene and none is older than the Cenozoic.^[9] although fossilized soils are preserved from as far back as the Archean.^[10]

Soil science has two basic branches of study: edaphology and pedology. Edaphology is concerned with the influence of soils on living things.^[11] Pedology is focused on the formation, description (morphology), and classification of soils in their natural environment.^[12] In engineering terms, soil is included in the broader concept of regolith, which also includes other loose material that lies above the bedrock.^[13] Soil is commonly referred to as *earth* or *dirt*; technically, the term *dirt* should be restricted to displaced soil!^[14]



A, B, and C represent the soil profile, a notation firstly coined by Vasily Dokuchaev, the father of pedology; A is the topsoil; B is a regolith; C is a saprolite, a less-weathered regolith; the bottom-most layer represents the bedrock.



Surface-water-gley developed in glacial till, Northern Ireland

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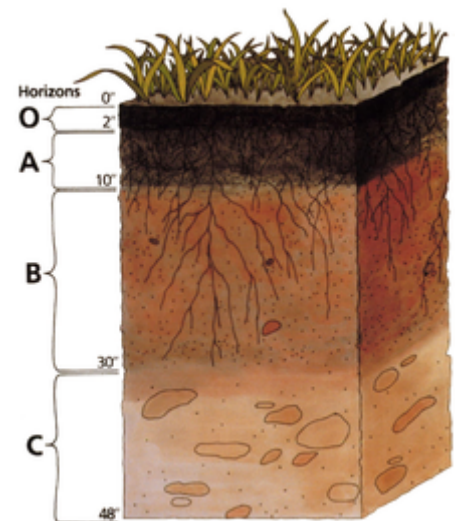
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Overview

Functions of soils

Soil is a major component of the Earth's ecosystem. The world's ecosystems are impacted in far-reaching ways by the processes carried out in the soil, from ozone depletion and global warming, to rainforest destruction and water pollution. With respect to Earth's carbon cycle, soil is an important carbon reservoir, and it is potentially one of the most reactive to human disturbance^[15] and climate change.^[16] As the planet warms, it has been predicted that soils will add carbon dioxide to the atmosphere due to increased biological activity at higher temperatures, a positive feedback (amplification).^[17] This prediction has, however, been questioned on consideration of more recent knowledge on soil carbon turnover.^[18]

Soil acts as an engineering medium, a habitat for soil organisms, a recycling system for nutrients and organic wastes, a regulator of water quality, a modifier of atmospheric composition, and a medium for plant growth, making it a critically important provider of ecosystem services.^[19] Since soil has a tremendous range of available niches and habitats, it contains most of the Earth's genetic diversity. A gram of soil can contain billions of organisms, belonging to thousands of species, mostly microbial and in the main still unexplored.^{[20][21]} Soil has a mean prokaryotic density of roughly 10^8 organisms per gram,^[22] whereas the ocean has no more than 10^7 prokaryotic organisms per milliliter (gram) of seawater.^[23] Organic carbon held in soil is eventually returned to the atmosphere through the process of respiration carried out by heterotrophic organisms, but a substantial part is retained in the soil in the form of soil organic matter; tillage usually increases the rate of soil respiration, leading to the depletion of soil organic matter.^[24] Since plant roots need oxygen, ventilation is an important characteristic of soil. This ventilation can be accomplished via networks of interconnected soil pores, which also absorb and hold rainwater making it readily available for uptake by plants. Since plants require a nearly continuous supply of water, but most regions receive sporadic rainfall, the water-holding capacity of soils is vital for plant survival.^[25]



Soil Profile: Darkened topsoil and reddish subsoil layers are typical in some regions.

Soils can effectively remove impurities,^[26] kill disease agents,^[27] and degrade contaminants, this latter property being called natural attenuation.^[28] Typically, soils maintain a net absorption of oxygen and methane, and undergo a net release of carbon dioxide and nitrous oxide.^[29] Soils offer plants physical support, air, water, temperature moderation, nutrients, and protection from toxins.^[30] Soils provide readily available nutrients to plants and animals by converting dead ganic matter into various nutrient forms.^[31]

Description

A typical soil is about 50% solids (45% mineral and 5% organic matter), and 50% voids of which half is occupied by water and half by gas.^[32] The percent soil mineral and organic content can be treated as a constant (in the short term), while the percent soil water and gas content is considered highly variable whereby a rise in one is simultaneously balanced by a reduction in the other.^[33] The pore space allows for the infiltration and movement of air and water, both of which are critical for life in soil.^[34] Compaction, a common problem with soils, reduces this space, preventing air and water from reaching plant roots and soil organisms.^[35]

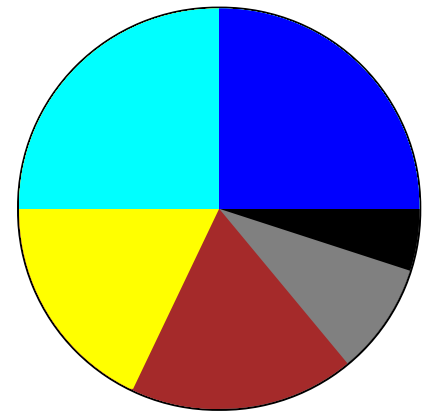
Given sufficient time, an undifferentiated soil will evolve a soil profile which consists of two or more layers, referred to as soil horizons, that differ in one or more properties such as in their texture, structure, density, porosity, consistency, temperature, color, and reactivity.^[9] The horizons differ greatly in thickness and generally lack sharp boundaries; their development is dependent on the type of parent material, the processes that modify those parent materials, and the soil-forming factors that influence those processes. The biological influences on soil properties are strongest near the surface, while the geochemical influences on soil properties increase with depth. Mature soil profiles typically include three basic master horizons: A, B and C. The solum normally includes the A and B horizons. The living component of the soil is largely confined to the solum, and is generally more prominent in the A horizon.^[36]

The soil texture is determined by the relative proportions of the individual particles of sand, silt, and clay that make up the soil. The interaction of the individual mineral particles with organic matter, water, gases via biotic and abiotic processes causes those particles to flocculate (stick together) to form aggregates or peds.^[37] Where these aggregates can be identified, a soil can be said to be developed, and can be described further in terms of color, porosity, consistency, reaction (acidity), etc.

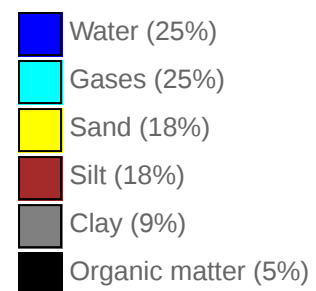
Water is a critical agent in soil development due to its involvement in the dissolution, precipitation, erosion, transport, and deposition of the materials of which a soil is composed.^[38] The mixture of water and dissolved or suspended materials that occupy the soil pore space is called the soil solution. Since soil water is never pure water, but contains hundreds of dissolved organic and mineral substances, it may be more accurately called the soil solution. Water is central to the dissolution, precipitation and leaching of minerals from the soil profile. Finally, water affects the type of vegetation that grows in a soil, which in turn affects the development of the soil, a complex feedback which is exemplified in the dynamics of banded vegetation patterns in semi-arid regions.^[39]

Soils supply plants with nutrients, most of which are held in place by particles of clay and organic matter (colloids).^[40] The nutrients may be adsorbed on clay mineral surfaces, bound within clay minerals (absorbed), or bound within organic compounds as part of the living organisms or dead soil organic matter. These bound nutrients interact with soil water to buffer the soil solution composition (attenuate changes in the soil solution) as soils wet up or dry out, as plants take up nutrients, as salts are leached, or as acids or alkalis are added.^{[41][42]}

Plant nutrient availability is affected by soil pH, which is a measure of the hydrogen ion activity in the soil solution. Soil pH is a function of many soil forming factors, and is generally lower (more acid) where weathering is more advanced.^[43]



Components of a loam soil by percent volume



Most plant nutrients, with the exception of nitrogen, originate from the minerals that make up the soil parent material. Some nitrogen originates from rain as dilute nitric acid and ammonia,^[44] but most of the nitrogen is available in soils as a result of nitrogen fixation by bacteria. Once in the soil-plant system, most nutrients are recycled through living organisms, plant and microbial residues (soil organic matter), mineral-bound forms, and the soil solution. Both living microorganisms and soil organic matter are of critical importance to this recycling, and thereby to soil formation and soil fertility.^[45] Microbial activity in soils may release nutrients from minerals or organic matter for use by plants and other microorganisms, sequester (incorporate) them into living cells, or cause their loss from the soil by volatilisation (loss to the atmosphere as gases) or leaching.

History of the study of soil

Studies of soil fertility

The history of the study of soil is intimately tied to our urgent need to provide food for ourselves and forage for our animals. Throughout history, civilizations have prospered or declined as a function of the availability and productivity of their soils.^[46]

The Greek historian Xenophon (450–355 BCE) is credited with being the first to expound upon the merits of green-manuring crops: "But then whatever weeds are upon the ground, being turned into earth, enrich the soil as much as dung."^[47]

Columella's "Husbandry," circa 60 CE, advocated the use of lime and that clover and alfalfa (green manure) should be turned under, and was used by 15 generations (450 years) under the Roman Empire until its collapse.^{[47][48]} From the fall of Rome to the French Revolution, knowledge of soil and agriculture was passed on from parent to child and as a result, crop yields were low. During the European Dark Ages, Yahya Ibn al-'Awvam's handbook,^[49] with its emphasis on irrigation, guided the people of North Africa, Spain and the Middle East; a translation of this work was finally carried to the southwest of the United States when under Spanish influence.^[50] Olivier de Serres, considered as the father of French agronomy, was the first to suggest the abandonment of fallowing and its replacement by hay meadows within crop rotations, and he highlighted the importance of soil (the French terroir) in the management of vineyards. His famous book *Le Théâtre d'Agriculture et mesnage des champs*^[51] contributed to the rise of modern, sustainable agriculture and to the collapse of old agricultural practices such as the lifting of forest litter for the amendment of crops (the French soutrage) and assarting, which ruined the soils of western Europe during Middle Ages and even later on according to regions.^[52]

Experiments into what made plants grow first led to the idea that the ash left behind when plant matter was burned was the essential element but overlooked the role of nitrogen, which is not left on the ground after combustion, a belief which prevailed until the 19th century.^[53] In about 1635, the Flemish chemist Jan Baptist van Helmont thought he had proved water to be the essential element from his famous five years' experiment with a willow tree grown with only the addition of rainwater. His conclusion came from the fact that the increase in the plant's weight had apparently been produced only by the addition of water, with no reduction in the soil's weight.^{[54][55]} John Woodward (d. 1728) experimented with various types of water ranging from clean to muddy and found muddy water the best, and so he concluded that earthy matter was the essential element. Others concluded it was humus in the soil that passed some essence to the growing plant. Still others held that the vital growth principal was something passed from dead plants or animals to the new plants. At the start of the 18th century, Jethro Tull demonstrated that it was beneficial to cultivate (stir) the soil, but his opinion that the stirring made the fine parts of soil available for plant absorption was erroneous.^{[54][56]}

As chemistry developed, it was applied to the investigation of soil fertility. The French chemist Antoine Lavoisier showed in about 1778 that plants and animals must [combust] oxygen internally to live and was able to deduce that most of the 165-pound weight of van Helmont's willow tree derived from air.^[57] It was the French agriculturalist Jean-Baptiste Boussingault who by means of experimentation obtained evidence showing that the main sources of carbon, hydrogen and oxygen for plants were air and water, while nitrogen was taken from soil.^[58] Justus von Liebig in his book *Organic chemistry in its applications to agriculture and physiology* (published 1840), asserted that the chemicals in plants must have come from the soil and air and that to maintain soil fertility, the used minerals must be replaced.^[59] Liebig nevertheless believed the nitrogen was supplied from the air. The enrichment of soil with guano by the Incas was rediscovered in 1802, by Alexander von Humboldt. This led to its mining and that of Chilean nitrate and to its application to soil in the United States and Europe after 1840.^[60]

The work of Liebig was a revolution for agriculture, and so other investigators started experimentation based on it. In England John Bennet Lawes and Joseph Henry Gilbert worked in the Rothamsted Experimental Station, founded by the former, and (re)discovered that plants took nitrogen from the soil, and that salts needed to be in an available state to be absorbed by plants. Their investigations also produced the "superphosphate", consisting in the acid treatment of phosphate rock.^[61] This led to the invention and use of salts of potassium (K) and nitrogen (N) as fertilizers. Ammonia generated by the production of coke was recovered and used as fertiliser.^[62] Finally, the chemical basis of nutrients delivered to the soil in manure was understood and in the mid-19th century chemical fertilisers were applied. However the dynamic interaction of soil and its life forms still awaited discovery.

In 1856 J. Thomas Way discovered that ammonia contained in fertilisers was transformed into nitrates,^[63] and twenty years later Robert Warington proved that this transformation was done by living organisms.^[64] In 1890 Sergei Winogradsky announced he had found the bacteria responsible for this transformation.^[65]

It was known that certain legumes could take up nitrogen from the air and fix it to the soil but it took the development of bacteriology towards the end of the 19th century to lead to an understanding of the role played in nitrogen fixation by bacteria. The symbiosis of bacteria and leguminous roots, and the fixation of nitrogen by the bacteria, were simultaneously discovered by the German agronomist Hermann Hellriegel and the Dutch microbiologist Martinus Beijerinck.^[61]

Crop rotation, mechanisation, chemical and natural fertilisers led to a doubling of wheat yields in western Europe between 1800 and 1900.^[66]

Studies of soil formation

The scientists who studied the soil in connection with agricultural practices had considered it mainly as a static substrate. However, soil is the result of evolution from more ancient geological materials, under the action of biotic and abiotic (not associated with life) processes. After studies of the improvement of the soil commenced, others began to study soil genesis and as a result also soil types and classifications.

In 1860, in Mississippi, Eugene W. Hilgard studied the relationship among rock material, climate, and vegetation, and the type of soils that were developed. He realised that the soils were dynamic, and considered soil types classification.^[67] Unfortunately his work was not continued. At the same time Vasily Dokuchaev (about 1870) was leading a team of soil scientists in Russia who conducted an extensive survey of soils, finding that similar basic rocks, climate and vegetation types lead to similar soil layering and types, and established the concepts for soil classifications. Due to language barriers, the work of this team was not communicated to western Europe until 1914 through a publication in German by Konstantin Dmitrievich Glinka, a member of the Russian team.^[68]

Curtis F. Marbut was influenced by the work of the Russian team, translated Glinka's publication into English,^[69] and as he was placed in charge of the U. S. National Cooperative Soil Survey applied it to a national soil classification system.^[54]

Soil formation

Soil formation, or pedogenesis, is the combined effect of physical, chemical, biological and anthropogenic processes working on soil parent material. Soil is said to be formed when organic matter has accumulated and colloids are washed downward, leaving deposits of clay, humus, iron oxide, carbonate, and gypsum, producing a distinct layer called the B horizon. This is a somewhat arbitrary definition as mixtures of sand, silt, clay and humus will support biological and agricultural activity before that time. These constituents are moved from one level to another by water and animal activity. As a result, layers (horizons) form in the soil profile. The alteration and movement of materials within a soil causes the formation of distinctive soil horizons. However, more recent definitions of soil embrace soils without any organic matter, such as those regoliths that formed on Mars^[70] and analogous conditions in planet Earth deserts.^[71]

An example of the development of a soil would begin with the weathering of lava flow bedrock, which would produce the purely mineral-based parent material from which the soil texture forms. Soil development would proceed most rapidly from bare rock of recent flows in a warm climate, under heavy and frequent rainfall. Under such conditions, plants (in a first stage nitrogen-fixing lichens and cyanobacteria then epilithic higher plants) become established very quickly on basaltic lava, even though there is very

little organic material. The plants are supported by the porous rock as it is filled with nutrient-bearing water that carries minerals dissolved from the rocks. Crevasses and pockets, local topography of the rocks, would hold fine materials and harbour plant roots. The developing plant roots are associated with mineral-weathering mycorrhizal fungi^[72] that assist in breaking up the porous lava, and by these means organic matter and a finer mineral soil accumulate with time. Such initial stages of soil development have been described on volcanoes,^[73] inselbergs,^[74] and glacial moraines.^[75]

Soil-forming factors

How soil formation proceeds is influenced by at least five classic factors that are intertwined in the evolution of a soil. They are: parent material, climate, topography (relief), organisms, and time.^[76] When reordered to climate, relief, organisms, parent material, and time, they form the acronym CROPT^[77]

Parent material

The mineral material from which a soil forms is called parent material. Rock, whether its origin is igneous, sedimentary, or metamorphic, is the source of all soil mineral materials and the origin of all plant nutrients with the exceptions of nitrogen, hydrogen and carbon. As the parent material is chemically and physically weathered, transported, deposited and precipitated, it is transformed into a soil.

Typical soil parent mineral materials are:^[78]

- Quartz: SiO_2
- Calcite: CaCO_3
- Feldspar: KAlSi_3O_8
- Mica (biotite): $\text{K}(\text{Mg,Fe})_3\text{AlSi}_3\text{O}_{10}(\text{OH})_2$

Classification of parent material

Parent materials are classified according to how they came to be deposited. Residual materials are mineral materials that have weathered in place from primary bedrock. Transported materials are those that have been deposited by water, wind, ice or gravity. Cumulose material is organic matter that has grown and accumulates in place.

Residual soils are soils that develop from their underlying parent rocks and have the same general chemistry as those rocks. The soils found on mesas, plateaux, and plains are residual soils. In the United States as little as three percent of the soils are residual.^[79]

Most soils derive from transported materials that have been moved many miles by wind, water, ice and gravity.



Loess field in Germany

- Aeolian processes (movement by wind) are capable of moving silt and fine sand many hundreds of miles, forming loess soils (60–90 percent silt);^[80] common in the Midwest of North America, north-western Europe, Argentina and Central Asia. Clay is seldom moved by wind as it forms stable aggregates.
- Water-transported materials are classed as either alluvial, lacustrine, or marine. Alluvial materials are those moved and deposited by flowing water. Sedimentary deposits settled in lakes are called lacustrine. Lake Bonneville and many soils around the Great Lakes of the United States are examples. Marine deposits, such as soils along the Atlantic and Gulf Coasts and in the Imperial Valley of California of the United States, are the beds of ancient seas that have been revealed as the land uplifted.
- Ice moves parent material and makes deposits in the form of terminal and lateral moraines in the case of stationary glaciers. Retreating glaciers leave smoother ground moraines and in all cases, outwash plains are left as alluvial deposits are moved downstream from the glacier.
- Parent material moved by gravity is obvious at the base of steep slopes as talus cones and is called colluvial material.

Cumulose parent material is not moved but originates from deposited organic material. This includes peat and muck soils and results from preservation of plant residues by the low oxygen content of a high water table. While peat may form sterile soils, muck soils may be very fertile.

Weathering of parent material

The weathering of parent material takes the form of physical weathering (disintegration), chemical weathering (decomposition) and chemical transformation. Generally, minerals that are formed under high temperatures and pressures at great depths within the Earth's mantle are less resistant to weathering, while minerals formed at low temperature and pressure environment of the surface are more resistant to weathering. Weathering is usually confined to the top few meters of geologic material, because physical, chemical, and biological stresses and fluctuations generally decrease with depth.^[81] Physical disintegration begins as rocks that have solidified deep in the Earth are exposed to lower pressure near the surface and swell and become mechanically unstable. Chemical decomposition is a function of mineral solubility, the rate of which doubles with each 10 °C rise in temperature, but is strongly dependent on water to effect chemical changes. Rocks that will decompose in a few years in tropical climates will remain unaltered for millennia in deserts.^[6] Structural changes are the result of hydration, oxidation, and reduction. Chemical weathering mainly results from the excretion of organic acids and chelating compounds by bacteria^[82] and fungi,^[83] thought to increase under present-day greenhouse effect.^[84]

- **Physical disintegration** is the first stage in the transformation of parent material into soil. Temperature fluctuations cause expansion and contraction of the rock, splitting it along lines of weakness. Water may then enter the cracks and freeze and cause the physical splitting of material along a path toward the center of the rock, while temperature gradients within the rock can cause exfoliation of "shells". Cycles of wetting and drying cause soil particles to be abraded to a finer size, as does the physical rubbing of material as it is moved by wind, water and gravity. Water can deposit within rocks minerals that expand upon drying, thereby stressing the rock. Finally, organisms reduce parent material in size and create crevices and pores through the mechanical action of plant roots and the digging activity of animals.^[85] Grinding of parent material by rock-eating animals also contributes to incipient soil formation.^[86]
- **Chemical decomposition and structural changes** result when minerals are made soluble by water or are changed in structure. The first three of the following list are solubility changes and the last three are structural changes.^[87]
 1. The **solution** of salts in water results from the action of bipolar water molecules on ionic salt compounds producing a solution of ions and water removing those minerals and reducing the rock's integrity at a rate depending on water flow and pore channels.^[88]
 2. **Hydrolysis** is the transformation of minerals into polar molecules by the splitting of intervening water. This results in soluble acid-base pairs. For example, the hydrolysis of orthoclase feldspar transforms it to acid silicate clay and basic potassium hydroxide, both of which are more soluble.^[89]
 3. In **carbonation**, the solution of carbon dioxide in water forms carbonic acid. Carbonic acid will transform calcite into more soluble calcium bicarbonate.^[90]
 4. **Hydration** is the inclusion of water in a mineral structure, causing it to swell and leaving it stressed and easily decomposed.^[91]
 5. **Oxidation** of a mineral compound is the inclusion of oxygen in a mineral, causing it to increase its oxidation number and swell due to the relatively large size of oxygen, leaving it stressed and more easily attacked by water (hydrolysis) or carbonic acid (carbonation).^[92]
 6. **Reduction**, the opposite of oxidation, means the removal of oxygen, hence the oxidation number of some part of the mineral is reduced, which occurs when oxygen is scarce. The reduction of minerals leaves them electrically unstable, more soluble and internally stressed and easily decomposed. It mainly occurs in waterlogged conditions.^[93]

Of the above, hydrolysis and carbonation are the most effective, in particular in regions of high rainfall, temperature and physical erosion.^[94] Chemical weathering becomes more effective as the surface area of the rock increases, thus is favoured by physical disintegration.^[95] This stems in latitudinal and altitudinal climate gradients in regolith formation.^{[96][97]}

Saprolite is a particular example of a residual soil formed from the transformation of granite, metamorphic and other types of bedrock into clay minerals. Often called [weathered granite], saprolite is the result of weathering processes that include: hydrolysis, chelation from organic compounds, hydration (the solution of minerals in water with resulting cation and anion pairs) and physical processes that include freezing and thawing. The mineralogical and chemical composition of the primary bedrock material, its physical features, including grain size and degree of consolidation, and the rate and type of weathering transforms the parent material into a different

mineral. The texture, pH and mineral constituents of saprolite are inherited from its parent material. This process is also called *arenization*, resulting in the formation of sandy soils (granitic arenas), thanks to the much higher resistance of quartz compared to other mineral components of granite (micas, amphiboles, feldspars).^[98]

Climate

The principal climatic variables influencing soil formation are effective precipitation (i.e., precipitation minus evapotranspiration) and temperature, both of which affect the rates of chemical, physical, and biological processes. Temperature and moisture both influence the organic matter content of soil through their effects on the balance between primary production and decomposition: the colder or drier the climate the lesser atmospheric carbon is fixed as organic matter while the lesser organic matter is decomposed.^[99]

Climate is the dominant factor in soil formation, and soils show the distinctive characteristics of the climate zones in which they form, with a feedback to climate through transfer of carbon stocked in soil horizons back to the atmosphere.^[16] If warm temperatures and abundant water are present in the profile at the same time, the processes of weathering, leaching, and plant growth will be maximized. According to the climatic determination of biomes, humid climates favor the growth of trees. In contrast, grasses are the dominant native vegetation in subhumid and semiarid regions, while shrubs and brush of various kinds dominate in arid areas.^[100]

Water is essential for all the major chemical weathering reactions. To be effective in soil formation, water must penetrate the regolith. The seasonal rainfall distribution, evaporative losses, site topography, and soil permeability interact to determine how effectively precipitation can influence soil formation. The greater the depth of water penetration, the greater the depth of weathering of the soil and its development. Surplus water percolating through the soil profile transports soluble and suspended materials from the upper layers (eluviation) to the lower layers (illuviation), including clay particles^[101] and dissolved organic matter.^[102] It may also carry away soluble materials in the surface drainage waters. Thus, percolating water stimulates weathering reactions and helps differentiate soil horizons. Likewise, a deficiency of water is a major factor in determining the characteristics of soils of dry regions. Soluble salts are not leached from these soils, and in some cases they build up to levels that curtail plant^[103] and microbial growth.^[104] Soil profiles in arid and semi-arid regions are also apt to accumulate carbonates and certain types of expansive clays (calcrete or caliche horizons).^{[105][106]} In tropical soils, when the soil has been deprived of vegetation (e.g. by deforestation) and thereby is submitted to intense evaporation, the upward capillary movement of water, which has dissolved iron and aluminum salts, is responsible for the formation of a superficial hard pan of laterite or bauxite, respectively, which is improper for cultivation, a known case of irreversible soil degradation (lateritization, bauxitization).^[107]

The direct influences of climate include.^[108]

- A shallow accumulation of lime in low rainfall areas as caliche
- Formation of acid soils in humid areas
- Erosion of soils on steep hillsides
- Deposition of eroded materials downstream
- Very intense chemical weathering, leaching and erosion in warm and humid regions where soil does not freeze

Climate directly affects the rate of weathering and leaching. Wind moves sand and smaller particles (dust), especially in arid regions where there is little plant cover, depositing it close^[109] or far from the entrainment source.^[110] The type and amount of precipitation influence soil formation by affecting the movement of ions and particles through the soil, and aid in the development of different soil profiles. Soil profiles are more distinct in wet and cool climates, where organic materials may accumulate, than in wet and warm climates, where organic materials are rapidly consumed.^[111] The effectiveness of water in weathering parent rock material depends on seasonal and daily temperature fluctuations, which favour tensile stresses in rock minerals, and thus their mechanical disaggregation, a process called *thermal fatigue*.^[112] By the same process freeze-thaw cycles are an effective mechanism which breaks up rocks and other consolidated materials.^[113]

Climate also indirectly influences soil formation through the effects of vegetation cover and biological activity, which modify the rates of chemical reactions in the soil.^[114]

Topography

The topography, or relief, is characterized by the inclination (slope), elevation, and orientation of the terrain. Topography determines the rate of precipitation or runoff and rate of formation or erosion of the surface soil profile. The topographical setting may either hasten or retard the work of climatic forces.

Steep slopes encourage rapid soil loss by erosion and allow less rainfall to enter the soil before running off and hence, little mineral deposition in lower profiles. In semiarid regions, the lower effective rainfall on steeper slopes also results in less complete vegetative cover, so there is less plant contribution to soil formation. For all of these reasons, steep slopes prevent the formation of soil from getting very far ahead of soil destruction. Therefore, soils on steep terrain tend to have rather shallow, poorly developed profiles in comparison to soils on nearby more level sites.^[115]

In swales and depressions where runoff water tends to concentrate, the regolith is usually more deeply weathered and soil profile development is more advanced. However, in the lowest landscape positions, water may saturate the regolith to such a degree that drainage and aeration are restricted. Here, the weathering of some minerals and the decomposition of organic matter are retarded, while the loss of iron and manganese is accelerated. In such low-lying topography, special profile features characteristic of wetland soils may develop. Depressions allow the accumulation of water, minerals and organic matter and in the extreme, the resulting soils will be saline marshes or peat bogs. Intermediate topography affords the best conditions for the formation of an agriculturally productive soil.

Organisms

Soil is the most abundant ecosystem on Earth, but the vast majority of organisms in soil are microbes, a great many of which have not been described.^{[116][117]} There may be a population limit of around one billion cells per gram of soil, but estimates of the number of species vary widely from 50,000 per gram to over a million per gram of soil.^{[116][118]} The total number of organisms and species can vary widely according to soil type, location, and depth.^{[117][118]}

Plants, animals, fungi, bacteria and humans affect soil formation (see soil biomantle and stonelayer). Soil animals, including soil macrofauna and soil mesofauna, mix soils as they form burrows and pores, allowing moisture and gases to move about, a process called bioturbation.^[119] In the same way, plant roots penetrate soil horizons and open channels upon decomposition.^[120] Plants with deep taproots can penetrate many metres through the different soil layers to bring up nutrients from deeper in the profile.^[121] Plants have fine roots that excrete organic compounds (sugars, organic acids, mucigel), slough off cells (in particular at their tip) and are easily decomposed, adding organic matter to soil, a process called rhizodeposition.^[122] Micro-organisms, including fungi and bacteria, effect chemical exchanges between roots and soil and act as a reserve of nutrients in a soil biological hotspot called rhizosphere.^[123] The growth of roots through the soil stimulates microbial populations, stimulating in turn the activity of their predators (notably amoeba), thereby increasing the mineralization rate, and in last turn root growth, a positive feedback called the soil microbial loop.^[124] Out of root influence, in the bulk soil, most bacteria are in a quiescent stage, forming microaggregates, i.e. mucilaginous colonies to which clay particles are glued, offering them a protection against desiccation and predation by soil microfauna (bacteriophagous protozoa and nematodes).^[125] Microaggregates (20-250 µm) are ingested by soil mesofauna and macrofauna, and bacterial bodies are partly or totally digested in their guts.^[126]

Humans impact soil formation by removing vegetation cover with erosion, waterlogging, lateritization or podzolization (according to climate and topography) as the result.^[127] Their tillage also mixes the different soil layers, restarting the soil formation process as less weathered material is mixed with the more developed upper layers, resulting in net increased rate of mineral weathering.^[128]

Earthworms, ants, termites, moles, gophers, as well as some millipedes and tenebrionid beetles mix the soil as they burrow, significantly affecting soil formation.^[129] Earthworms ingest soil particles and organic residues, enhancing the availability of plant nutrients in the material that passes through their bodies.^[130] They aerate and stir the soil and create stable soil aggregates, after having disrupted links between soil particles during the intestinal transit of ingested soil,^[131] thereby assuring ready infiltration of water.^[132] In addition, as ants and termites build mounds, they transport soil materials from one horizon to another.^[133] Other important functions are fulfilled by earthworms in the soil ecosystem, in particular their intense mucus production, both within the intestine and as a lining in their galleries,^[134] exert a priming effect on soil microflora,^[135] giving them the status of ecosystem engineers, which they share with ants and termites.^[136]

In general, the mixing of the soil by the activities of animals, sometimes called pedoturbation, tends to undo or counteract the tendency of other soil-forming processes that create distinct horizons.^[137] Termites and ants may also retard soil profile development by denuding large areas of soil around their nests, leading to increased loss of soil by erosion.^[138] Large animals such as gophers, moles, and prairie dogs bore into the lower soil horizons, bringing materials to the surface.^[139] Their tunnels are often open to the surface, encouraging the movement of water and air into the subsurface layers. In localized areas, they enhance mixing of the lower and upper horizons by creating, and later refilling, underground tunnels. Old animal burrows in the lower horizons often become filled with soil material from the overlying A horizon, creating profile features known as crotovinas.^[140]

Vegetation impacts soils in numerous ways. It can prevent erosion caused by excessive rain that might result from surface runoff.^[141] Plants shade soils, keeping them cooler^[142] and slow evaporation of soil moisture,^[143] or conversely, by way of transpiration, plants can cause soils to lose moisture, resulting in complex and highly variable relationships between leaf area index (measuring light interception) and moisture loss: more generally plants prevent soil from desiccation during driest months while they dry it during moister months, thereby acting as a buffer against strong moisture variation.^[144] Plants can form new chemicals that can break down minerals, both directly^[145] and indirectly through mycorrhizal fungi^[83] and rhizosphere bacteria,^[146] and improve the soil structure.^[147] The type and amount of vegetation depends on climate, topography, soil characteristics and biological factors, mediated or not by human activities.^{[148][149]} Soil factors such as density, depth, chemistry, pH, temperature and moisture greatly affect the type of plants that can grow in a given location. Dead plants and fallen leaves and stems begin their decomposition on the surface. There, organisms feed on them and mix the organic material with the upper soil layers; these added organic compounds become part of the soil formation process.^[150]

Human activities widely influence soil formation.^[151] For example, it is believed that Native Americans regularly set fires to maintain several large areas of prairie grasslands in Indiana and Michigan, although climate and mammalian grazers (e.g. bisons) are also advocated to explain the maintenance of the Great Plains of North America.^[152] In more recent times, human destruction of natural vegetation and subsequent tillage of the soil for crop production has abruptly modified soil formation.^[153] Likewise, irrigating soil in an arid region drastically influences soil-forming factors,^[154] as does adding fertilizer and lime to soils of low fertility.^[155]

Time

Time is a factor in the interactions of all the above.^[76] While a mixture of sand, silt and clay constitute the texture of a soil and the aggregation of those components produces peds, the development of a distinct B horizon marks the development of a soil or pedogenesis.^[156] With time, soils will evolve features that depend on the interplay of the prior listed soil-forming factors.^[76] It takes decades^[157] to several thousand years for a soil to develop a profile,^[158] although the notion of soil development has been criticized, soil being in a constant state-of-change under the influence of fluctuating soil-forming factors.^[159] That time period depends strongly on climate, parent material, relief, and biotic activity.^{[160][161]} For example, recently deposited material from a flood exhibits no soil development as there has not been enough time for the material to form a structure that further defines soil.^[162] The original soil surface is buried, and the formation process must begin anew for this deposit. Over time the soil will develop a profile that depends on the intensities of biota and climate. While a soil can achieve relative stability of its properties for extended periods,^[158] the soil life cycle ultimately ends in soil conditions that leave it vulnerable to erosion.^[163] Despite the inevitability of soil retrogression and degradation, most soil cycles are long.^[158]

Soil-forming factors continue to affect soils during their existence, even on "stable" landscapes that are long-enduring, some for millions of years.^[158] Materials are deposited on top^[164] or are blown or washed from the surface.^[165] With additions, removals and alterations, soils are always subject to new conditions. Whether these are slow or rapid changes depends on climate, topography and biological activity.^[166]

Physical properties of soils

The physical properties of soils, in order of decreasing importance for ecosystem services such as crop production, are texture, structure, bulk density, porosity, consistency, temperature, colour and resistivity.^[167] Soil texture is determined by the relative proportion of the three kinds of soil mineral particles, called soil separates: sand, silt, and clay. At the next larger scale, soil structures

called peds or more commonly *soil aggregates* are created from the soil separates when iron oxides, carbonates, clay, silica and humus, coat particles and cause them to adhere into larger, relatively stable secondary structures.^[168] Soil bulk density, when determined at standardized moisture conditions, is an estimate of soil compaction.^[169] Soil porosity consists of the void part of the soil volume and is occupied by gases or water. Soil consistency is the ability of soil materials to stick together. Soil temperature and colour are self-defining. Resistivity refers to the resistance to conduction of electric currents and affects the rate of corrosion of metal and concrete structures which are buried in soil.^[170] It also helps to estimate soil moisture.^[171] These properties vary through the depth of a soil profile, i.e. through soil horizons. Most of these properties determine the aeration of the soil and the ability of water to infiltrate and to be held within the soil.^[172]

Influence of Soil Texture Separates on Some Properties of Soils^[54]

Property/behavior	Sand	Silt	Clay
Water-holding capacity	Low	Medium to high	High
Aeration	Good	Medium	Poor
Drainage rate	High	Slow to medium	Very slow
Soil organic matter level	Low	Medium to high	High to medium
Decomposition of organic matter	Rapid	Medium	Slow
Warm-up in spring	Rapid	Moderate	Slow
Compactability	Low	Medium	High
Susceptibility to wind erosion	Moderate (High if fine sand)	High	Low
Susceptibility to water erosion	Low (unless fine sand)	High	Low if aggregated, otherwise high
Shrink/Swell Potential	Very Low	Low	Moderate to very high
Sealing of ponds, dams, and landfills	Poor	Poor	Good
Suitability for tillage after rain	Good	Medium	Poor
Pollutant leaching potential	High	Medium	Low (unless cracked)
Ability to store plant nutrients	Poor	Medium to High	High
Resistance to pH change	Low	Medium	High

Texture

The mineral components of soil are sand, silt and clay, and their relative proportions determine a soil's texture. Properties that are influenced by soil texture include porosity, permeability, infiltration, shrink-swell rate, water-holding capacity, and susceptibility to erosion. In the illustrated USDA textural classification triangle, the only soil in which neither sand, silt nor clay predominates is called loam. While even pure sand, silt or clay may be considered a soil, from the perspective of conventional agriculture a loam soil with a small amount of organic material is considered "ideal", inasmuch as fertilizers or manure are currently used to mitigate nutrient losses due to crop yields in the long term.^[173] The mineral constituents of a loam soil might be 40% sand, 40% silt and the balance 20% clay by weight. Soil texture affects soil behaviour, in particular its retention capacity for nutrients (e.g., cation exchange capacity)^[174] and water.

Sand and silt are the products of physical and chemical weathering of the parent rock.^[76] clay, on the other hand, is most often the product of the precipitation of the dissolved parent rock as a secondary mineral, except when derived from the weathering of mica.^[175] It is the surface area to volume ratio (specific surface area) of soil particles and the unbalanced ionic electric charges within those that determine their role in the fertility of soil, as measured by its cation exchange capacity.^{[176][177]} Sand is least active, having the least specific surface area, followed by silt; clay is the most active. Sand's greatest benefit to soil is that it resists

compaction and increases soil porosity, although this property stands only for pure sand, not for sand mixed with smaller minerals which fill the voids among sand grains.^[178] Silt is mineralogically like sand but with its higher specific surface area it is more chemically and physically active than sand. But it is the clay content of soil, with its very high specific surface area and generally large number of negative charges, that gives a soil its high retention capacity for water and nutrients.^[176] Clay soils also resist wind and water erosion better than silty and sandy soils, as the particles bond tightly to each other,^[179] and that with a strong mitigation effect of organic matter.^[180]

Sand is the most stable of the mineral components of soil; it consists of rock fragments, primarily quartz particles, ranging in size from 2.0 to 0.05 mm (0.0787 to 0.0020 in) in diameter. Silt ranges in size from 0.05 to 0.002 mm (0.002 to 0.00008 in). Clay cannot be resolved by optical microscopes as its particles are 0.002 mm (7.9×10^{-5} in) or less in diameter and a thickness of only 10 angstroms (10^{-10} m).^{[181][182]} In medium-textured soils, clay is often washed downward through the soil profile (a process called eluviation) and accumulates in the subsoil (a process called illuviation). There is no clear relationship between the size of soil mineral components and their mineralogical nature: sand and silt particles can be calcareous as well as siliceous,^[183] while textural clay (0.002 mm (7.9×10^{-5} in)) can be made of very fine quartz particles as well as of multi-layered secondary minerals.^[184] Soil mineral components belonging to a given textural class may thus share properties linked to their specific surface area (e.g. moisture retention) but not those linked to their chemical composition (e.g. cation exchange capacity).

Soil components larger than 2.0 mm (0.079 in) are classed as rock and gravel and are removed before determining the percentages of the remaining components and the textural class of the soil, but are included in the name. For example, a sandy loam soil with 20% gravel would be called gravelly sandy loam.

When the organic component of a soil is substantial, the soil is called organic soil rather than mineral soil. A soil is called organic if:

1. Mineral fraction is 0% clay and organic matter is 20% or more
2. Mineral fraction is 0% to 50% clay and organic matter is between 20% and 30%
3. Mineral fraction is 50% or more clay and organic matter 30% or more.^[185]

Structure

The clumping of the soil textural components of sand, silt and clay causes aggregates to form and the further association of those aggregates into larger units creates soil structures called peds (a contraction of the word pedolith). The adhesion of the soil textural components by organic substances, iron oxides, carbonates, clays, and silica, the breakage of those aggregates from expansion-contraction caused by freezing-thawing and wetting-drying cycles,^[186] and the build-up of aggregates by soil animals, microbial colonies and root tips^[187] shape soil into distinct geometric forms.^{[37][129]} The peds evolve into units which have various shapes, sizes and degrees of development.^[188] A soil clod, however, is not a ped but rather a mass of soil that results from mechanical disturbance of the soil such as cultivation. Soil structure affects aeration, water movement, conduction of heat, plant root growth and resistance to erosion.^[189] Water, in turn, has a strong effect on soil structure, directly via the dissolution and precipitation of minerals, the mechanical destruction of aggregates (slaking)^[190] and indirectly by promoting plant, animal and microbial growth.



Soil types by clay, silt and sand composition as used by the USDA



Iron-rich soil near Paint Pots in Kootenay National Park, Canada

Soil structure often gives clues to its texture, organic matter content, biological activity, past soil evolution, human use, and the chemical and mineralogical conditions under which the soil formed. While texture is defined by the mineral component of a soil and is an innate property of the soil that does not change with agricultural activities, soil structure can be improved or destroyed by the choice and timing of farming practices.^[37]

Soil structural classes:^[191]

1. Types: **Shape** and arrangement of peds

1. Platy: Peds are flattened one atop the other 1–10 mm thick. Found in the A-horizon of forest soils and lake sedimentation.
2. Prismatic and Columnar: Prismlike peds are long in the vertical dimension, 10–100 mm wide. Prismatic peds have flat tops, columnar peds have rounded tops. Tend to form in the B-horizon in high sodium soil where clay has accumulated.
3. Angular and subangular: Blocky peds are imperfect cubes, 5–50 mm, angular have sharp edges, subangular have rounded edges. Tend to form in the B-horizon where clay has accumulated and indicate poor water penetration.
4. Granular and Crumb: Spheroid peds of polyhedrons, 1–10 mm, often found in the A-horizon in the presence of organic material. Crumb peds are more porous and are considered ideal.

2. Classes: **Size** of peds whose ranges depend upon the above type

1. Very fine or very thin: <1 mm platy and spherical; <5 mm blocky; <10 mm prismlike.
2. Fine or thin: 1–2 mm platy and spherical; 5–10 mm blocky; 10–20 mm prismlike.
3. Medium: 2–5 mm platy granular; 10–20 mm blocky; 20–50 mm prismlike.
4. Coarse or thick: 5–10 mm platy granular; 20–50 mm blocky; 50–100 mm prismlike.
5. Very coarse or very thick: >10 mm platy granular; >50 mm blocky; >100 mm prismlike.

3. Grades: Is a measure of the degree of **development** or cementation within the peds that results in their strength and stability.

1. Weak: Weak cementation allows peds to fall apart into the three textural constituents, sand, silt and clay.
2. Moderate: Peds are not distinct in undisturbed soil but when removed they break into aggregates, some broken aggregates and little unaggregated material. This is considered ideal.
3. Strong: Peds are distinct before removed from the profile and do not break apart easily.
4. Structureless: Soil is entirely cemented together in one great mass such as slabs of clay or no cementation at all such as with sand.

At the largest scale, the forces that shape a soil's structure result from swelling and shrinkage that initially tend to act horizontally, causing vertically oriented prismatic peds. This mechanical process is mainly exemplified in the development of vertisols.^[192] Clayey soil, due to its differential drying rate with respect to the surface, will induce horizontal cracks, reducing columns to blocky peds.^[193] Roots, rodents, worms, and freezing-thawing cycles further break the peds into smaller peds of a more or less spherical shape.^[187]

At a smaller scale, plant roots extend into voids (macropores) and remove water^[194] causing macroporosity to increase and microporosity to decrease,^[195] thereby decreasing aggregate size.^[196] At the same time, root hairs and fungal hyphae create microscopic tunnels that break up peds.^{[197][198]}

At an even smaller scale, soil aggregation continues as bacteria and fungi exude sticky polysaccharides which bind soil into smaller peds.^[199] The addition of the raw organic matter that bacteria and fungi feed upon encourages the formation of this desirable soil structure.^[200]

At the lowest scale, the soil chemistry affects the aggregation or dispersal of soil particles. The clay particles contain polyvalent cations which give the faces of clay layers localized negative charges.^[201] At the same time, the edges of the clay plates have a slight positive charge, thereby allowing the edges to adhere to the negative charges on the faces of other clay particles or to flocculate (form clumps).^[202] On the other hand, when monovalent ions, such as sodium, invade and displace the polyvalent cations, they weaken the positive charges on the edges, while the negative surface charges are relatively strengthened. This leaves negative charge on the clay faces that repel other clay causing the particles to push apart, and by doing so deflocculate clay suspensions.^[203] As a result, the clay

disperses and settles into voids between peds, causing those to close. In this way the open structure of the soil is destroyed and the soil is made impenetrable to air and water.^[204] Such sodic soil (also called haline soil) tends to form columnar peds near the surface.^[205]

Density

Soil particle density is typically 2.60 to 2.75 grams per cm³ and is usually unchanging for a given soil.^[8] Soil particle density is lower for soils with high organic matter content,^[206] and is higher for soils with high iron-oxides content.^[207] Soil bulk density is equal to the dry mass of the soil divided by the volume of the soil; i.e., it includes air space and organic materials of the soil volume. Thereby soil bulk density is always less than soil particle density and is a good indicator of soil compaction.^[208] The soil bulk density of cultivated loam is about 1.1 to 1.4 g/cm³ (for comparison water is 1.0 g/cm³).^[209] Contrary to particle density, soil bulk density is highly variable for a given soil, with a strong causal relationship with soil biological activity and management strategies.^[210] However, it has been shown that, depending on species and the size of their aggregates (faeces), earthworms may either increase or decrease soil bulk density.^[211] A lower bulk density by itself does not indicate suitability for plant growth due to the confounding influence of soil texture and structure.^[212] A high bulk density is indicative of either soil compaction or a mixture of soil textural classes in which small particles fill the voids among coarser particles.^[213] Hence the positive correlation between the fractal dimension of soil, considered as a porous medium, and its bulk density,^[214] that explains the poor hydraulic conductivity of silty clay loam in the absence of a faunal structure.^[215]

Representative bulk densities of soils. The percentage pore space was calculated using 2.7 g/cm³ for particle density except for the peat soil, which is estimated.^[216]

Soil treatment and identification	Bulk density g/cm ³	Pore space %
Tilled surface soil of a cotton field	1.3	51
Trafficked inter-rows where wheels passed surface	1.67	37
Traffic pan at 25 cm deep	1.7	36
Undisturbed soil below traffic pan, clay loam	1.5	43
Rocky silt loam soil under aspen forest	1.62	40
Loamy sand surface soil	1.5	43
Decomposed peat	0.55	65

Porosity

Pore space is that part of the bulk volume of soil that is not occupied by either mineral or organic matter but is open space occupied by either gases or water. In a productive, medium-textured soil the total pore space is typically about 50% of the soil volume.^[217] Pore size varies considerably; the smallest pores (cryptopores; <0.1 μm) hold water too tightly for use by plant roots; plant-available water is held in ultramicropores, micropores and mesopores (0.1–75 μm); and macropores (>75 μm) are generally air-filled when the soil is at field capacity.

Soil texture determines total volume of the smallest pores;^[218] clay soils have smaller pores, but more total pore space than sands,^[219] despite of a much lower permeability.^[220] Soil structure has a strong influence on the larger pores that affect soil aeration, water infiltration and drainage.^[221] Tillage has the short-term benefit of temporarily increasing the number of pores of large size, but these can be rapidly degraded by the destruction of soil aggregation.^[222]

The pore size distribution affects the ability of plants and other organisms to access water and oxygen; large, continuous pores allow rapid transmission of air, water and dissolved nutrients through soil, and small pores store water between rainfall or irrigation events.^[223] Pore size variation also compartmentalizes the soil pore space such that many microbial and faunal organisms are not in direct competition with one another, which may explain not only the large number of species present, but the fact that functionally redundant organisms (organisms with the same ecological niche) can co-exist within the same soil.^[224]

Consistency

Consistency is the ability of soil to stick to itself or to other objects (cohesion and adhesion, respectively) and its ability to resist deformation and rupture. It is of approximate use in predicting cultivation problems^[225] and the engineering of foundations.^[226] Consistency is measured at three moisture conditions: air-dry, moist, and wet.^[227] In those conditions the consistency quality depends upon the clay content. In the wet state, the two qualities of stickiness and plasticity are assessed. A soil's resistance to fragmentation and crumbling is assessed in the dry state by rubbing the sample. Its resistance to shearing forces is assessed in the moist state by thumb and finger pressure. Additionally, the cemented consistency depends on cementation by substances other than clay, such as calcium carbonate, silica, oxides and salts; moisture content has little effect on its assessment. The measures of consistency border on subjective compared to other measures such as pH, since they employ the apparent feel of the soil in those states.

The terms used to describe the soil consistency in three moisture states and a last not affected by the amount of moisture are as follows:

1. Consistency of Dry Soil: loose, soft, slightly hard, hard, very hard, extremely hard
2. Consistency of Moist Soil: loose, very friable, friable, firm, very firm, extremely firm
3. Consistency of Wet Soil: nonsticky, slightly sticky, sticky, very sticky; nonplastic, slightly plastic, plastic, very plastic
4. Consistency of Cemented Soil: weakly cemented, strongly cemented, indurated (requires hammer blows to break up)^[228]

Soil consistency is useful in estimating the ability of soil to support buildings and roads. More precise measures of soil strength are often made prior to construction.

Temperature

Soil temperature depends on the ratio of the energy absorbed to that lost.^[229] Soil has a temperature range between -20 to 60 °C, with a mean annual temperature from -10 to 26 °C according to biomes.^[230] Soil temperature regulates seed germination,^[231] breaking of seed dormancy,^{[232][233]} plant and root growth^[234] and the availability of nutrients.^[235] Soil temperature has important seasonal, monthly and daily variations, fluctuations in soil temperature being much lower with increasing soil depth.^[236] Heavy mulching (a type of soil cover) can slow the warming of soil in summer and, at the same time, reduce fluctuations in surface temperature.^[237]

Most often, agricultural activities must adapt to soil temperatures by:

1. maximizing germination and growth by timing of planting (also determined by photoperiod)^[238]
2. optimizing use of anhydrous ammonia by applying to soil below 10 °C (50 °F)^[239]
3. preventing heaving and thawing due to frosts from damaging shallow-rooted crops^[240]
4. preventing damage to desirable soil structure by freezing of saturated soils^[241]
5. improving uptake of phosphorus by plants^[242]

Soil temperatures can be raised by drying soils^[243] or the use of clear plastic mulches.^[244] Organic mulches slow the warming of the soil.^[237]

There are various factors that affect soil temperature, such as water content,^[245] soil color,^[246] and relief (slope, orientation, and elevation),^[247] and soil cover (shading and insulation), in addition to air temperature.^[248] The color of the ground cover and its insulating properties have a strong influence on soil temperature.^[249] Whiter soil tends to have a higher albedo than blacker soil cover, which encourages whiter soils to have lower soil temperatures.^[246] The specific heat of soil is the energy required to raise the temperature of soil by 1 °C. The specific heat of soil increases as water content increases, since the heat capacity of water is greater than that of dry soil.^[250] The specific heat of pure water is ~ 1 calorie per gram, the specific heat of dry soil is ~ 0.2 calories per gram, hence, the specific heat of wet soil is ~ 0.2 to 1 calories per gram (0.8 to 4.2 kJ per kilogram).^[251] Also, a tremendous energy (~540 cal/g or 2260 kJ/kg) is required to evaporate water (known as the heat of vaporization). As such, wet soil usually warms more slowly than dry soil – wet surface soil is typically 3 to 6 °C colder than dry surface soil.^[252]

Soil heat flux refers to the rate at which heat energy moves through the soil in response to a temperature difference between two points in the soil. The heat flux density is the amount of energy that flows through soil per unit area per unit time and has both magnitude and direction. For the simple case of conduction into or out of the soil in the vertical direction, which is most often applicable the heat flux density is:

$$q_x = -k \frac{\delta T}{\delta x}$$

In SI units

q is the heat flux density, in SI the units are $\text{W}\cdot\text{m}^{-2}$

k is the soils' conductivity, $\text{W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$. The thermal conductivity is sometimes a constant, otherwise an average value of conductivity for the soil condition between the surface and the point at depth is used.

δT is the temperature difference (temperature gradient) between the two points in the soil between which the heat flux density is to be calculated. In SI the units are kelvin, K.

δx is the distance between the two points within the soil, at which the temperatures are measured and between which the heat flux density is being calculated. In SI the units are meters m, and where x is measured positive downward.

Heat flux is in the direction opposite the temperature gradient, hence the minus sign. That is to say if the temperature of the surface is higher than at depth x the negative sign will result in a positive value for the heat flux q, and which is interpreted as the heat being conducted into the soil.

[54]

Component	Thermal Conductivity ($\text{W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$)
Quartz	8.8
Clay	2.9
Organic matter	0.25
Water	0.57
Ice	2.4
Air	0.025
Dry soil	0.2-0.4
Wet soil	1–3

Soil temperature is important for the survival and early growth of seedlings.^[253] Soil temperatures affect the anatomical and morphological character of root systems.^[254] All physical, chemical, and biological processes in soil and roots are affected in particular because of the increased viscosities of water and protoplasm at low temperatures.^[255] In general, climates that do not preclude survival and growth of white spruce above ground are sufficiently benign to provide soil temperatures able to maintain white spruce root systems. In some northwestern parts of the range, white spruce occurs on permafrost sites^[256] and although young unignified roots of conifers may have little resistance to freezing,^[257] the root system of containerized white spruce was not affected by exposure to a temperature of less than 30 °C.^[258]

Optimum temperatures for tree root growth range between 10 °C and 25 °C in general^[259] and for spruce in particular.^[260] In 2-week-old white spruce seedlings that were then grown for 6 weeks in soil at temperatures of 15 °C, 19 °C, 23 °C, 27 °C, and 31 °C; shoot height, shoot dry weight, stem diameter, root penetration, root volume, and root dry weight all reached maxima at 19 °C.^[261]

However, whereas strong positive relationships between soil temperature (5 °C to 25 °C) and growth have been found in trembling aspen and balsam poplar, white and other spruce species have shown little or no changes in growth with increasing soil temperature.^{[260][262][263][264][265]} Such insensitivity to soil low temperature may be common among a number of western and

boreal conifers.^[266]

Color

Soil colour is often the first impression one has when viewing soil. Striking colours and contrasting patterns are especially noticeable. The Red River of the South carries sediment eroded from extensive reddish soils like Port Silt Loam in Oklahoma. The Yellow River in China carries yellow sediment from eroding loess soils. Mollisols in the Great Plains of North America are darkened and enriched by organic matter. Podisols in boreal forests have highly contrasting layers due to acidity and leaching.

In general, color is determined by the organic matter content, drainage conditions, and degree of oxidation. Soil color, while easily discerned, has little use in predicting soil characteristics.^[267] It is of use in distinguishing boundaries of horizons within a soil profile,^[268] determining the origin of a soil's parent material,^[269] as an indication of wetness and waterlogged conditions,^[270] and as a qualitative means of measuring organic,^[271] iron oxide^[272] and clay contents of soils.^[273] Color is recorded in the Munsell color system as for instance 10YR3/4 *Dusky Red*, with 10YR as *hue*, 3 as *value* and 4 as *chroma*. Munsell color dimensions (hue, value and chroma) can be averaged among samples and treated as quantitative parameters, displaying significant correlations with various soil^[274] and vegetation properties.^[275]

Soil color is primarily influenced by soil mineralogy. Many soil colours are due to various iron minerals.^[272] The development and distribution of colour in a soil profile result from chemical and biological weathering, especially redox reactions.^[270] As the primary minerals in soil parent material weather, the elements combine into new and colourful compounds. Iron forms secondary minerals of a yellow or red colour,^[276] organic matter decomposes into black and brown humic compounds,^[277] and manganese^[278] and sulfur^[279] can form black mineral deposits. These pigments can produce various colour patterns within a soil. Aerobic conditions produce uniform or gradual colour changes, while reducing environments (anaerobic) result in rapid colour flow with complex, mottled patterns and points of colour concentration.^[280]

Resistivity

Soil resistivity is a measure of a soil's ability to retard the conduction of an electric current. The electrical resistivity of soil can affect the rate of galvanic corrosion of metallic structures in contact with the soil. Higher moisture content or increased electrolyte concentration can lower resistivity and increase conductivity, thereby increasing the rate of corrosion.^{[281][282]} Soil resistivity values typically range from about 2 to 1000 Ω -m, but more extreme values are not unusual.^[283]

Soil water

Water that enters a field is removed from a field by runoff, drainage, evaporation or transpiration. Runoff is the water that flows on the surface to the edge of the field; drainage is the water that flows through the soil downward or toward the edge of the field underground; evaporative water loss from a field is that part of the water that evaporates into the atmosphere directly from the field's surface; transpiration is the loss of water from the field by its evaporation from the plant itself.

Water affects soil formation, structure, stability and erosion but is of primary concern with respect to plant growth. Water is essential to plants for four reasons:

1. It constitutes 80%-95% of the plant's protoplasm.
2. It is essential for photosynthesis.
3. It is the solvent in which nutrients are carried to, into and throughout the plant.
4. It provides the turgidity by which the plant keeps itself in proper position.^[284]

In addition, water alters the soil profile by dissolving and re-depositing minerals, often at lower levels, and possibly leaving the soil sterile in the case of extreme rainfall and drainage. In a loam soil, solids constitute half the volume, gas one-quarter of the volume, and water one-quarter of the volume of which only half of which will be available to most plants.

A flooded field will drain the gravitational water under the influence of gravity until water's adhesive and cohesive forces resist further drainage at which point it is said to have reached field capacity.^[285] At that point, plants must apply suction^{[285][286]} to draw water from a soil. The water that plants may draw from the soil is called the available water.^{[285][287]} Once the available water is used up the remaining moisture is called unavailable water as the plant cannot produce sufficient suction to draw that water in. A plant must produce suction that increases from zero for a flooded field to 1/3 bar at field dry condition (one bar is a little less than one atmosphere pressure). At 15 bar suction, wilting percent, seeds will not germinate,^{[288][285][289]} plants begin to wilt and then die. Water moves in soil under the influence of gravity, osmosis and capillarity. When water enters the soil, it displaces air from some of the pores, since air content of a soil is inversely related to its water content.

The rate at which a soil can absorb water depends on the soil and its other conditions. As a plant grows, its roots remove water from the largest pores first. Soon the larger pores hold only air, and the remaining water is found only in the intermediate- and smallest-sized pores. The water in the smallest pores is so strongly held to particle surfaces that plant roots cannot pull it away. Consequently, not all soil water is available to plants. When saturated, the soil may lose nutrients as the water drains. Water moves in a draining field under the influence of pressure where the soil is locally saturated and by capillarity pull to dryer parts of the soil. Most plant water needs are supplied from the suction caused by evaporation from plant leaves and 10% is supplied by "suction" created by osmotic pressure differences between the plant interior and the soil water. Plant roots must seek out water. Insufficient water will damage the yield of a crop. Most of the available water is used in transpiration to pull nutrients into the plant.

Water retention forces

Water is retained in a soil when the adhesive force of attraction that water's hydrogen atoms have for the oxygen of soil particles is stronger than the cohesive forces that water's hydrogen feels for other water oxygen atoms.^[290] When a field is flooded, the soil pore space is completely filled by water. The field will drain under the force of gravity until it reaches what is called field capacity, at which point the smallest pores are filled with water and the largest with water and gases.^[291] The total amount of water held when field capacity is reached is a function of the specific surface area of the soil particles. As a result, high clay and high organic soils have higher field capacities. The total force required to pull or push water out of soil is termed suction and usually expressed in units of bars (10^5 pascal) which is just a little less than one-atmosphere pressure. Alternatively, the terms "tension" or "moisture potential" may be used.^[292]

Moisture classification

The forces with which water is held in soils determine its availability to plants. Forces of adhesion hold water strongly to mineral and humus surfaces and less strongly to itself by cohesive forces. A plant's root may penetrate a very small volume of water that is adhering to soil and be initially able to draw in water that is only lightly held by the cohesive forces. But as the droplet is drawn down, the forces of adhesion of the water for the soil particles make reducing the st produce increasingly higher suction, finally up to 15 bar. At 15 bar suction, the soil water amount is called wilting percent. At that suction the plant cannot sustain its water needs as water is still being lost from the plant by transpiration; the plant's turgidity is lost, and it wilts. The next level, called air-dry, occurs at 1000 bar suction. Finally the oven dry condition is reached at 10,000 bar suction. All water below wilting percentage is called unavailable water.^[293]

Soil moisture content

When the soil moisture content is optimal for plant growth, the water in the large and intermediate size pores can move about in the soil and be easily used by plants. The amount of water remaining in a soil drained to field capacity and the amount that is available are functions of the soil type. Sandy soil will retain very little water, while clay will hold the maximum amount. The time required to drain a field from flooded condition for a clay loam that begins at 43% water by weight to a field capacity of 22% is six days, whereas a sand loam that is flooded to its maximum of 22% water will take two days to reach field capacity of 11% water. The available water for the clay loam might be 1% whereas for the sand loam it might be only 8% by weight.^[294]

Wilting point, field capacity and available water capacity of various soil textures^[295]

Soil Texture	Wilting Point		Field Capacity		Available water capacity	
	Water per foot of soil depth		Water per foot of soil depth		Water per foot of soil depth	
	%	in.	%	in.	%	in.
Medium sand	1.7	0.3	6.8	1.2	5.1	0.9
Fine sand	2.3	0.4	8.5	1.5	6.2	1.1
Sandy loam	3.4	0.6	11.3	2.0	7.9	1.4
Fine sandy loam	4.5	0.8	14.7	2.6	10.2	1.8
Loam	6.8	1.2	18.1	3.2	11.3	2.0
Silt loam	7.9	1.4	19.8	3.5	11.9	2.1
Clay loam	10.2	1.8	21.5	3.8	11.3	2.0
Clay	14.7	2.6	22.6	4.0	7.9	1.4

The above are average values for the soil textures as the percentages of sand, silt and clay vary

Water flow in soils

Water moves through soil due to the force of gravity, osmosis and capillarity. At zero to one-third bar suction, water is pushed through soil from the point of its application under the force of gravity and the pressure gradient created by the pressure of the water; this is called saturated flow. At higher suction, water movement is pulled by capillarity from wetter toward dryer soil. This is caused by water's adhesion to soil solids, and is called unsaturated flow^[296]

Water infiltration and movement in soil is controlled by six factors:

1. Soil texture
2. Soil structure. Fine-textured soils with granular structure are most favourable to infiltration of water
3. The amount of organic matter Coarse matter is best and if on the surface helps prevent the destruction of soil structure and the creation of crusts.
4. Depth of soil to impervious layers such as hardpans or bedrock
5. The amount of water already in the soil
6. Soil temperature. Warm soils take in water faster while frozen soils may not be able to absorb depending on the type of freezing.^[297]

Water infiltration rates range from 0.25 cm (0.098 in) per hour for high clay soils to 2.5 cm (0.98 in) per hour for sand and well stabilised and aggregated soil structures.^[298] Water flows through the ground unevenly, called "gravity fingers", because of the surface tension between water particles.^[299] Tree roots create paths for rainwater flow through soil by breaking through soil including clay layers: one study showed roots increasing infiltration of water by 153% and another study showed an increase by 27 times.^[300] Flooding temporarily increases soil permeability in river beds, helping to recharge aquifers.^[301]

Saturated flow

Water applied to a soil is pushed by pressure gradients from the point of its application where it is saturated locally, to less saturated areas. Once soil is completely wetted, any more water will move downward, or percolate, carrying with it clay, humus and nutrients, primarily cations, out of the range of plant roots. In order of decreasing solubility, the leached nutrients are:

- Calcium
- Magnesium, Sulfur, Potassium; depending upon soil composition
- Nitrogen; usually little, unless nitrate fertiliser was applied recently
- Phosphorus; very little as its forms in soil are of low solubility^[302]

In the United States percolation water due to rainfall ranges from zero inches just east of the Rocky Mountains to twenty or more inches in the Appalachian Mountains and the north coast of the Gulf of Mexico.^[295]

Unsaturated flow

At suctions less than one-third bar, water moves in all directions via unsaturated flow at a rate that is dependent on the square of the diameter of the water-filled pores. Water is pulled by capillary action due to the adhesion force of water to the soil solids, producing a suction gradient from wet towards drier soil. Doubling the diameter of the pores increases the flow rate by a factor of four. Large pores drained by gravity and not filled with water do not greatly increase the flow rate for unsaturated flow. Water flow is primarily from coarse-textured soil into fine-textured soil and is slowest in fine-textured soils such as clay^[303]

Water uptake by plants

Of equal importance to the storage and movement of water in soil is the means by which plants acquire it and their nutrients. Ninety percent of water is taken up by plants as passive absorption caused by the pulling force of water evaporating (transpiring) from the long column of water that leads from the plant's roots to its leaves. In addition, the high concentration of salts within plant roots creates an osmotic pressure gradient that pushes soil water into the roots. Osmotic absorption becomes more important during times of low water transpiration caused by lower temperatures (for example at night) or high humidity. It is the process that causes guttation.^[304]

Root extension is vital for plant survival. A study of a single winter rye plant grown for four months in one cubic foot of loam soil showed that the plant developed 13,800,000 roots, a total of 385 miles in length with 2,550 square feet in surface area; and 14 billion hair roots of 6,600 miles total length and 4,320 square feet total area; for a total surface area of 6,870 square feet (83 ft squared). The total surface area of the loam soil was estimated to be 560,000 square feet.^[305] In other words, the roots were in contact with only 1.2% of the soil.

Roots must seek out water as the unsaturated flow of water in soil can move only at a rate of up to 2.5 cm (one inch) per day; as a result they are constantly dying and growing as they seek out high concentrations of soil moisture. Insufficient soil moisture, to the point of causing wilting, will cause permanent damage and crop yields will suffer. When grain sorghum was exposed to soil suction as low as 13.0 bar during the seed head emergence through bloom and seed set stages of growth, its production was reduced by 34%.^[306]

Consumptive use and water efficiency

Only a small fraction (0.1% to 1%) of the water used by a plant is held within the plant. The majority is ultimately lost via transpiration, while evaporation from the soil surface is also substantial. Transpiration plus evaporative soil moisture loss is called evapotranspiration. Evapotranspiration plus water held in the plant totals to consumptive use, which is nearly identical to evapotranspiration.^[306]

The total water used in an agricultural field includes runoff, drainage and consumptive use. The use of loose mulches will reduce evaporative losses for a period after a field is irrigated, but in the end the total evaporative loss will approach that of an uncovered soil. The benefit from mulch is to keep the moisture available during the seedling stage. Water use efficiency is measured by transpiration ratio, which is the ratio of the total water transpired by a plant to the dry weight of the harvested plant. Transpiration ratios for crops range from 300 to 700. For example, alfalfa may have a transpiration ratio of 500 and as a result 500 kilograms of water will produce one kilogram of dry alfalfa.^[307]

Soil atmosphere

The atmosphere of soil is radically different from the atmosphere above. The consumption of oxygen, by microbes and plant roots and their release of carbon dioxide, decrease oxygen and increase carbon dioxide concentration. Atmospheric CO₂ concentration is 0.04%, but in the soil pore space it may range from 10 to 100 times that level. At extreme levels CO₂ is toxic. In addition, the soil voids are saturated with water vapour. Adequate porosity is necessary, not just to allow the penetration of water, but also to allow gases to diffuse in and out. Movement of gases is by diffusion from high concentrations to lower. Oxygen diffuses in and is consumed and excess levels of carbon dioxide, diffuse out with other gases as well as water. Soil texture and structure strongly affect soil porosity and gas diffusion. It is the total pore space (porosity) of soil not the pore size that determines the rate of diffusion of gases

into and out of soil.^[308] A Platy soil structure and compacted soils (low porosity) impede gas flow, and a deficiency of oxygen may encourage anaerobic bacteria to reduce (strip oxygen) from nitrate NO_3 to the gases N_2 , N_2O , and NO , which are then lost to the atmosphere, thereby depleting the soil of nitrogen. Aerated soil is also a net sink of methane CH_4 but a net producer of greenhouse gases when soils are depleted of oxygen and subject to elevated temperatures.^[309]

Composition of soil particles

Soil particles can be classified by their chemical composition (mineralogy) as well as their size. The particle size distribution of a soil, its texture, determines many of the properties of that soil, but the mineralogy of those particles can strongly modify those properties. The mineralogy of the finest soil particles, clays especially important.^[310]

Gravel, sand and silt

Gravel, sand and silt are the larger soil particles, and their mineralogy is often inherited from the parent material of the soil, but may include products of weathering (such as concretions of calcium carbonate or iron oxide), or residues of plant and animal life (such as silica phytoliths).^[311] Quartz is the most common mineral in the sand or silt fraction as it is resistant to chemical weathering; other common minerals are feldspars, micas and ferromagnesian minerals such as pyroxenes, amphiboles and olivines.^[311]

Mineral colloids; soil clays

Due to its high specific surface area and its unbalanced negative charges, clay is the most active mineral component of soil. It is a colloidal and most often a crystalline material. In soils, clay is a soil textural class and is defined in a physical sense as any mineral particle less than $2\text{ }\mu\text{m}$ ($8 \times 10^{-5}\text{ in}$) in effective diameter. Many soil minerals, such as gypsum, carbonates, or quartz, are small enough to be classified as clay based on their physical size, but chemically they do not afford the same utility as do clay minerals. Chemically, clay is a range of minerals with certain reactive properties.^[312]

Clay was once thought to be very small particles of quartz, feldspar, mica, hornblende or augite, but it is now known to be (with the exception of mica-based clays) a precipitate with a mineralogical composition that is dependent on but different from its parent materials and is classed as a secondary mineral. The type of clay that is formed is a function of the parent material and the composition of the minerals in solution. Clay minerals continue to be formed as long as the soil exists.^[313] Mica-based clays result from a modification of the primary mica mineral in such a way that it behaves and is classed as a clay. Most clays are crystalline, but some are amorphous. The clays of a soil are a mixture of the various types of clay, but one type predominates.^[314]

There are four groups of clay: layer silicates; crystalline chain silicates; metal oxides and hydroxides and oxy-oxides; and amorphous and allophanes. Most clays are crystalline and most are made up of three or four planes of oxygen held together by planes of aluminium and silicon by way of ionic bonds that together form a single layer of clay. The spatial arrangement of the oxygen atoms determines clay's structure. Half of the weight of clay is oxygen, but on a volume basis oxygen is ninety percent.^[315] The layers of clay are sometimes held together through hydrogen bonds or potassium bridges and as a result will swell less in the presence of water. Other clays, such as montmorillonite, have layers that are loosely attached and will swell greatly when water intervenes between the layers.^[316]

There are four groups of clays:

1. Layer Crystalline **alumino-silica clays** montmorillonite, illite, vermiculite, chlorite, kaolinite.
2. Crystalline Chain **carbonate and sulfate minerals** calcite (CaCO_3), dolomite ($\text{CaMg}(\text{CO}_3)_2$) and gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$).
3. **Amorphous clays** young mixtures of silica ($\text{SiO}_2\text{-OH}$) and alumina ($\text{Al}(\text{OH})_3$) which have not had time to form regular crystals.
4. **Sesquioxide clays** old, highly leached clays which result in oxides of iron, aluminium and titanium.^[317]

Alumino-silica clays

Alumino-silica clays are characterised by their regular crystalline structure. Oxygen in ionic bonds with silicon forms a tetrahedral coordination (silicon at the center) which in turn forms sheets of silica. Two sheets of silica are bonded together by a plane of aluminium which forms an octahedral coordination, called alumina, with the oxygens of the silica sheet above and that below it. Hydroxyl ions (OH^-) sometimes substitute for oxygen. During the clay formation process, Al^{3+} may substitute for Si^{4+} in the silica layer, and as much as one fourth of the aluminium Al^{3+} may be substituted by Zn^{2+} , Mg^{2+} or Fe^{2+} in the alumina layer. The substitution of lower-valence cations for higher-valence cations (isomorphous substitution) gives clay a local negative charge on an oxygen atom that attracts and holds water and positively charged soil cations, some of which are of value for plant growth. Isomorphous substitution occurs during the clay's formation and does not change with time.^{[318][160]}

- **Montmorillonite** clay is made of four planes of oxygen with two silicon and one central aluminium plane intervening. The alumino-silicate montmorillonite clay is said to have a 2:1 ratio of silicon to aluminium. The seven planes together form a single crystal of montmorillonite. The crystals are weakly held together and water may intervene, causing the clay to swell up to ten times its dry volume. It occurs in soils which have had little leaching, hence it is found in arid regions. As the crystals are not bonded face to face, the entire surface is exposed and available for surface reactions, hence it has a high cation exchange capacity (CEC).^{[319][320][321]}
- **Illite** is a 2:1 clay similar in structure to montmorillonite but has potassium bridges between the faces of the clay crystals and the degree of swelling depends on the degree of weathering of the potassium. The active surface area is reduced due to the potassium bonds. Illite originates from the modification of mica, a primary mineral. It is often found together with montmorillonite and its primary minerals. It has moderate CEC.^{[322][320][323][324][325]}
- **Vermiculite** is a mica-based clay similar to illite, but the crystals of clay are held together more loosely by hydrated magnesium and it will swell, but not as much as does montmorillonite. It has very high CEC.^{[326][327][324][325]}
- **Chlorite** is similar to vermiculite, but the loose bonding by occasional hydrated magnesium, as in vermiculite, is replaced by a hydrated magnesium sheet, that firmly bonds the planes above and below it. It has two planes of silicon, one of aluminium and one of magnesium; hence it is a 2:2 clay. Chlorite does not swell and it has low CEC.^{[326][328]}
- **Kaolinite** is very common, highly weathered clay and more common than montmorillonite in acid soils. It has one silica and one alumina plane per crystal; hence it is a 1:1 type clay. One plane of silica of montmorillonite is dissolved and is replaced with hydroxyls, which produces strong hydrogen bonds to the oxygen in the next crystal of clay. As a result, kaolinite does not swell in water and has a low specific surface area, and as almost no isomorphous substitution has occurred it has a low CEC. Where rainfall is high, acid soils selectively leach more silica than alumina from the original clays, leaving kaolinite. Even heavier weathering results in sesquioxide clays.^{[329][315][321][323][330][331]}

Crystalline chain clays

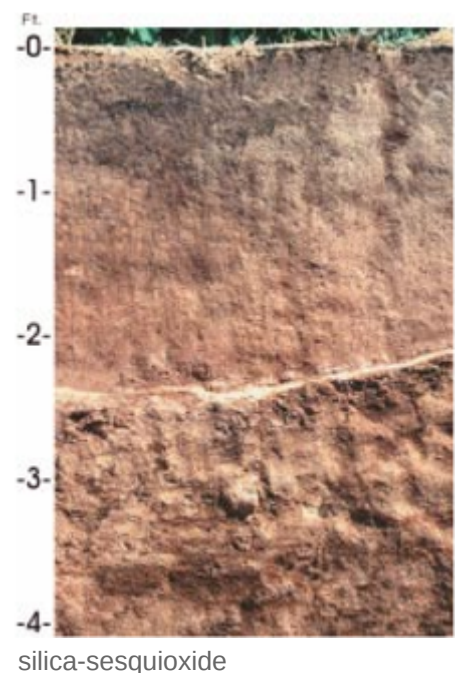
The carbonate and sulfate minerals are much more soluble and hence are found primarily in desert soils where leaching is less active.

Amorphous clays

Amorphous clays are young, and commonly found in volcanic ash. They are mixtures of alumina and silica which have not formed the ordered crystal shape of alumino-silica clays which time would provide. The majority of their negative charges originates from hydroxyl ions, which can gain or lose a hydrogen ion (H^+) in response to soil pH, in such way as to buffer the soil pH. They may have either a negative charge provided by the attached hydroxyl ion (OH^-), which can attract a cation, or lose the hydrogen of the hydroxyl to solution and display a positive charge which can attract anions. As a result, they may display either high CEC in an acid soil solution, or high anion exchange capacity in a basic soil solution.^[329]

Sesquioxide clays

Sesquioxide clays are a product of heavy rainfall that has leached most of the silica from alumino-silica clay, leaving the less soluble oxides iron hematite (Fe_2O_3), iron hydroxide ($\text{Fe}(\text{OH})_3$), aluminium hydroxide gibbsite ($\text{Al}(\text{OH})_3$), hydrated manganese birnessite (MnO_2). It takes hundreds of thousands of years of leaching to create sesquioxide clays. *Sesqui* is Latin for "one and one-half": there



are three parts oxygen to two parts iron or aluminium; hence the ratio is one and one-half (not true for all). They are hydrated and act as either amorphous or crystalline. They are not sticky and do not swell, and soils high in them behave much like sand and can rapidly pass water. They are able to hold large quantities of phosphates. Sesquioxides have low CEC but are able to hold anions as well as cations. Such soils range from yellow to red in colour. Such clays tend to hold phosphorus so tightly that it is unavailable for absorption by plants.^{[332][333][334]}

Organic colloids

Humus is the final state of decomposition of organic matter. While it may linger for a thousand years, on the larger scale of the age of the mineral soil components, it is temporary. It is composed of the very stable lignins (30%) and complex sugars (polyuronides, 30%), proteins (30%), waxes, and fats that are resistant to breakdown by microbes. Its chemical assay is 60% carbon, 5% nitrogen, some oxygen and the remainder hydrogen, sulfur, and phosphorus. On a dry weight basis, the CEC of humus is many times greater than that of clay.^{[335][336][337]}

Carbon and terra preta

In the extreme environment of high temperatures and the leaching caused by the heavy rain of tropical rain forests, the clay and organic colloids are largely destroyed. The heavy rains wash the alumino-silicate clays from the soil leaving only sesquioxide clays of low CEC. The high temperatures and humidity allow bacteria and fungi to virtually dissolve any organic matter on the rain-forest floor overnight and much of the nutrients are volatilized or leached from the soil and lost. However, carbon in the form of charcoal is far more stable than soil colloids and is capable of performing many of the functions of the soil colloids of sub-tropical soils. Soil containing substantial quantities of charcoal, of an anthropogenic origin, is called terra preta. Research into terra preta is still young but is promising. Fallow periods "on the Amazonian Dark Earths can be as short as 6 months, whereas fallow periods on oxisols are usually 8 to 10 years long"^[338]

Soil chemistry

The chemistry of a soil determines its ability to supply available plant nutrients and affects its physical properties and the health of its microbial population. In addition, a soil's chemistry also determines its corrosivity, stability, and ability to absorb pollutants and to filter water. It is the surface chemistry of mineral and organic colloids that determines soil's chemical properties. "A colloid is a small, insoluble, nondiffusible particle larger than a molecule but small enough to remain suspended in a fluid medium without settling. Most soils contain organic colloidal particles called humus as well as the inorganic colloidal particles of clays." The very high specific surface area of colloids and their net charges, gives soil its ability to hold and release ions. Negatively charged sites on colloids attract and release cations in what is referred to as cation exchange. Cation-exchange capacity (CEC) is the amount of exchangeable cations per unit weight of dry soil and is expressed in terms of milliequivalents of positively charged ions per 100 grams of soil (or centimoles of positive charge per kilogram of soil; cmol_c/kg). Similarly, positively charged sites on colloids can attract and release anions in the soil giving the soil anion exchange capacity (AEC).

Cation and anion exchange

The cation exchange, that takes place between colloids and soil water, buffers (moderates) soil pH, alters soil structure, and purifies percolating water by adsorbing cations of all types, both useful and harmful.

The negative or positive charges on colloid particles make them able to hold cations or anions, respectively, to their surfaces. The charges result from four sources:^[339]

1. Isomorphous substitution occurs in clay during its formation, when lower-valence cations substitute for higher-valence cations in the crystal structure. Substitutions in the outermost layers are more effective than for the innermost layers, as the charge strength drops off as the square of the distance. The net result is oxygen atoms with net negative charge and the ability to attract cations.
2. Edge-of-clay oxygen atoms are not in balance ionically as the tetrahedral and octahedral structures are incomplete.

3. Hydroxyls may substitute for oxygens of the silica layers. When the hydrogens of the clay hydroxyls are ionised into solution, they leave the oxygen with a negative charge.

4. Hydrogens of humus hydroxyl groups may be ionised into solution, leaving an oxygen with a negative charge.

Cations held to the negatively charged colloids resist being washed downward by water and out of reach of plants' roots, thereby preserving the fertility of soils in areas of moderate rainfall and low temperatures.

There is a hierarchy in the process of cation exchange on colloids, as they differ in the strength of adsorption by the colloid and hence their ability to replace one another. If present in equal amounts in the soil water solution:

Al^{3+} replaces H^+ replaces Ca^{2+} replaces Mg^{2+} replaces K^+ same as NH_4^+ replaces Na^+

If one cation is added in large amounts, it may replace the others by the sheer force of its numbers. This is called mass action. This is largely what occurs with the addition of fertiliser.

As the soil solution becomes more acidic (low pH, and an abundance of H^+), the other cations more weakly bound to colloids are pushed into solution as hydrogen ions occupy those sites. A low pH may cause hydrogen of hydroxyl groups to be pulled into solution, leaving charged sites on the colloid available to be occupied by other cations. This ionisation of hydroxyl groups on the surface of soil colloids creates what is described as pH-dependent charges. Unlike permanent charges developed by isomorphous substitution, pH-dependent charges are variable and increase with increasing pH.^[42] Freed cations can be made available to plants but are also prone to be leached from the soil, possibly making the soil less fertile. Plants are able to excrete H^+ into the soil and by that means, change the pH of the soil near the root and push cations off the colloids, thus making those available to the plant.

Cation exchange capacity (CEC)

Cation exchange capacity should be thought of as the soil's ability to remove cations from the soil water solution and sequester those to be exchanged later as the plant roots release hydrogen ions to the solution. CEC is the amount of exchangeable hydrogen cation (H^+) that will combine with 100 grams dry weight of soil and whose measure is one milliequivalents per 100 grams of soil (1 meq/100 g). Hydrogen ions have a single charge and one-thousandth of a gram of hydrogen ions per 100 grams dry soil gives a measure of one milliequivalent of hydrogen ion. Calcium, with an atomic weight 40 times that of hydrogen and with a valence of two, converts to $(40/2) \times 1$ milliequivalent = 20 milliequivalents of hydrogen ion per 100 grams of dry soil or 20 meq/100 g.^[340] The modern measure of CEC is expressed as centimoles of positive charge per kilogram (cmol/kg) of oven-dry soil.

Most of the soil's CEC occurs on clay and humus colloids, and the lack of those in hot, humid, wet climates, due to leaching and decomposition respectively explains the relative sterility of tropical soils. Live plant roots also have some CEC.

Cation exchange capacity for soils; soil textures; soil colloids^[341]

Soil	State	CEC meq/100 g
Charlotte fine sand	Florida	1.0
Ruston fine sandy loam	Texas	1.9
Gloucester loam	New Jersey	11.9
Grundy silt loam	Illinois	26.3
Gleason clay loam	California	31.6
Susquehanna clay loam	Alabama	34.3
Davie mucky fine sand	Florida	100.8
Sands	-----	1 - 5
Fine sandy loams	-----	5-10
Loams and silt loams	-----	5-15
Clay loams	-----	15-30
Clays	-----	over 30
Sesquioxides	-----	0-3
Kaolinite	-----	3-15
Illite	-----	25-40
Montmorillonite	-----	60-100
Vermiculite (similar to illite)	-----	80-150
Humus	-----	100-300

Anion exchange capacity (AEC)

Anion exchange capacity should be thought of as the soil's ability to remove anions from the soil water solution and sequester those for later exchange as the plant roots release carbonate anions to the soil water solution. Those colloids which have low CEC tend to have some AEC. Amorphous and sesquioxide clays have the highest AEC, followed by the iron oxides. Levels of AEC are much lower than for CEC. Phosphates tend to be held at anion exchange sites.

Iron and aluminum hydroxide clays are able to exchange their hydroxide anions (OH^-) for other anions. The order reflecting the strength of anion adhesion is as follows:



The amount of exchangeable anions is of a magnitude of tenths to a few milliequivalents per 100 g dry soil.^[341] As pH rises, there are relatively more hydroxyls, which will displace anions from the colloids and force them into solution and out of storage; hence AEC decreases with increasing pH (alkalinity).

Soil reaction (pH)

Soil reactivity is expressed in terms of pH and is a measure of the acidity or alkalinity of the soil. More precisely, it is a measure of hydrogen ion concentration in an aqueous solution and ranges in values from 0 to 14 (acidic to basic) but practically speaking for soils, pH ranges from 3.5 to 9.5, as pH values beyond those extremes are toxic to life forms.

Soil pH

At 25 °C an aqueous solution that has a pH of 3.5 has $10^{-3.5}$ moles H^+ (hydrogen ions) per litre of solution (and also $10^{-10.5}$ mole/litre OH^-). A pH of 7, defined as neutral, has 10^{-7} moles hydrogen ions per litre of solution and also 10^{-7} moles of OH^- per litre; since the two concentrations are equal, they are said to neutralise each other. A pH of 9.5 has $10^{-9.5}$ moles hydrogen ions per litre of solution (and also $10^{-2.5}$ mole per litre OH^-). A pH of 3.5 has one million times more hydrogen ions per litre than a solution with pH of 9.5 ($9.5 - 3.5 = 6$ or 10^6) and is more acidic.^[342]

The effect of pH on a soil is to remove from the soil or to make available certain ions. Soils with high acidity tend to have toxic amounts of aluminium and manganese. Plants which need calcium need moderate alkalinity, but most minerals are more soluble in acid soils. Soil organisms are hindered by high acidity, and most agricultural crops do best with mineral soils of pH 6.5 and organic soils of pH 5.5.^[343]

In high rainfall areas, soils tend to acidity as the basic cations are forced off the soil colloids by the mass action of hydrogen ions from the rain as those attach to the colloids. High rainfall rates can then wash the nutrients out, leaving the soil sterile. Once the colloids are saturated with H^+ , the addition of any more hydrogen ions or aluminum hydroxyl cations drives the pH even lower (more acidic) as the soil has been left with no buffering capacity. In areas of extreme rainfall and high temperatures, the clay and humus may be washed out, further reducing the buffering capacity of the soil. In low rainfall areas, unleached calcium pushes pH to 8.5 and with the addition of exchangeable sodium, soils may reach pH 10. Beyond a pH of 9, plant growth is reduced. High pH results in low micro-nutrient mobility but water-soluble chelates of those nutrients can correct the deficit. Sodium can be reduced by the addition of gypsum (calcium sulphate) as calcium adheres to clay more tightly than does sodium causing sodium to be pushed into the soil water solution where it can be washed out by an abundance of water.^[344]

Base saturation percentage

There are acid-forming cations (hydrogen and aluminium) and there are base-forming cations. The fraction of the base-forming cations that occupy positions on the soil colloids is called the base saturation percentage. If a soil has a CEC of 20 meq and 5 meq are aluminium and hydrogen cations (acid-forming), the remainder of positions on the colloids ($20 - 5 = 15$ meq) are assumed occupied by base-forming cations, so that the percentage base saturation is $15/20 \times 100\% = 75\%$ (the complement 25% is assumed acid-forming cations). When the soil pH is 7 (neutral), base saturation is 100 percent and there are no hydrogen ions stored on the colloids. Base saturation is almost in direct proportion to pH (increases with increasing pH). It is of use in calculating the amount of lime needed to neutralise an acid soil. The amount of lime needed to neutralize a soil must take account of the amount of acid forming ions on the colloids not just those in the soil water solution. The addition of enough lime to neutralize the soil water solution will be insufficient to change the pH, as the acid forming cations stored on the soil colloids will tend to restore the original pH condition as they are pushed off those colloids by the calcium of the added lime.^[345]

Buffering of soils

The resistance of soil to change in pH, as a result of the addition of acid or basic material, is a measure of the buffering capacity of a soil and (for a particular soil type) increases as the CEC increases. Hence, pure sand has almost no buffering ability, while soils high in colloids have high buffering capacity. Buffering occurs by cation exchange and neutralisation.

The addition of a small amount highly basic aqueous ammonia to a soil will cause the ammonium to displace hydrogen ions from the colloids, and the end product is water and colloiddally fixed ammonium, but little permanent change overall in soil pH.

The addition of a small amount of lime, $CaCO_3$, will displace hydrogen ions from the soil colloids, causing the fixation of calcium to colloids and the evolution of CO_2 and water, with little permanent change in soil pH.

The above are examples of the buffering of soil pH. The general principal is that an increase in a particular cation in the soil water solution will cause that cation to be fixed to colloids (buffered) and a decrease in solution of that cation will cause it to be withdrawn from the colloid and moved into solution (buffered). The degree of buffering is often related to the CEC of the soil; the greater the CEC, the greater the buffering capacity of the soil.^[346]

Nutrients

Sixteen elements or nutrients are essential for plant growth and reproduction. They are carbon C, hydrogen H, oxygen O, nitrogen N, phosphorus P, potassium K, sulfur S, calcium Ca, magnesium Mg, iron Fe, boron B, manganese Mn, copper Cu, zinc Zn, molybdenum Mo, nickel Ni and chlorine Cl.^{[347][348][349]} Nutrients required for plants to complete their life cycle are considered **essential nutrients**. Nutrients that enhance the growth of plants but are not necessary to complete the plant's life cycle are considered **non-essential**. With the exception of carbon, hydrogen and oxygen, which are supplied by carbon dioxide and water, and nitrogen, provided through nitrogen fixation,^[349] the nutrients derive originally from the mineral component of the soil.

Plant uptake of nutrients can only proceed when they are present in a plant-available form. In most situations, nutrients are absorbed in an ionic form from (or together with) soil water. Although minerals are the origin of most nutrients, and the bulk of most nutrient elements in the soil is held in crystalline form within primary and secondary minerals, they weather too slowly to support rapid plant growth. For example, The application of finely ground minerals, feldspar and apatite, to soil seldom provides the necessary amounts of potassium and phosphorus at a rate sufficient for good plant growth, as most of the nutrients remain bound in the crystals of those minerals.^[350]

The nutrients adsorbed onto the surfaces of clay colloids and soil organic matter provide a more accessible reservoir of many plant nutrients (e.g. K, Ca, Mg, P, Zn). As plants absorb the nutrients from the soil water, the soluble pool is replenished from the surface-bound pool. The decomposition of soil organic matter by microorganisms is another mechanism whereby the soluble pool of nutrients is replenished – this is important for the supply of plant-available N, S, and B from soil.^[351]

Gram for gram, the capacity of humus to hold nutrients and water is far greater than that of clay minerals. All in all, small amounts of humus may remarkably increase the soil's capacity to promote plant growth.^{[352][351]}

Plant nutrients, their chemical symbols, and the ionic forms common in soils and available for plant uptake^[353]

Element	Symbol	Ion or molecule
Carbon	C	CO ₂ (mostly through leaves)
Hydrogen	H	H ⁺ , HOH (water)
Oxygen	O	O ²⁻ , OH ⁻ , CO ₃ ²⁻ , SO ₄ ²⁻ , CO ₂
Phosphorus	P	H ₂ PO ₄ ⁻ , HPO ₄ ²⁻ (phosphates)
Potassium	K	K ⁺
Nitrogen	N	NH ₄ ⁺ , NO ₃ ⁻ (ammonium, nitrate)
Sulfur	S	SO ₄ ²⁻
Calcium	Ca	Ca ²⁺
Iron	Fe	Fe ²⁺ , Fe ³⁺ (ferrous, ferric)
Magnesium	Mg	Mg ²⁺
Boron	B	H ₃ BO ₃ , H ₂ BO ₃ ⁻ , B(OH) ₄ ⁻
Manganese	Mn	Mn ²⁺
Copper	Cu	Cu ²⁺
Zinc	Zn	Zn ²⁺
Molybdenum	Mo	MoO ₄ ²⁻ (molybdate)
Chlorine	Cl	Cl ⁻ (chloride)

Soil processes important for nutrient uptake

Nutrients in the soil are taken up by the plant through its roots. To be taken up by a plant, a nutrient element must be located near the root surface; however, the supply of nutrients in contact with the root is rapidly depleted. There are three basic mechanisms whereby nutrient ions dissolved in the soil solution are brought into contact with plant roots:

1. Mass flow of water
2. Diffusion within water
3. Interception by root growth

All three mechanisms operate simultaneously, but one mechanism or another may be most important for a particular nutrient. For example, in the case of calcium, which is generally plentiful in the soil solution, mass flow alone can usually bring sufficient amounts to the root surface. However, in the case of phosphorus, diffusion is needed to supplement mass flow. For the most part, nutrient ions must travel some distance in the soil solution to reach the root surface. This movement can take place by mass flow, as when dissolved nutrients are carried along with the soil water flowing toward a root that is actively drawing water from the soil. In this type of movement, the nutrient ions are somewhat analogous to leaves floating down a stream. In addition, nutrient ions continually move by diffusion from areas of greater concentration toward the nutrient-depleted areas of lower concentration around the root surface. That process is due to random motion of molecules. By this means, plants can continue to take up nutrients even at night, when water is only slowly absorbed into the roots as transpiration has almost stopped. Finally, root interception comes into play as roots continually grow into new undepleted soil.

Estimated relative importance of mass flow, diffusion and root interception as mechanisms in supplying plant nutrients to corn plant roots in soils^[354]

Nutrient	Approximate percentage supplied by:		
	Mass flow	Root interception	Diffusion
Nitrogen	98.8	1.2	0
Phosphorus	6.3	2.8	90.9
Potassium	20.0	2.3	77.7
Calcium	71.4	28.6	0
Sulfur	95.0	5.0	0
Molybdenum	95.2	4.8	0

In the above table, phosphorus and potassium nutrients move more by diffusion than they do by mass flow in the soil water solution, as they are rapidly taken up by the roots creating a concentration of almost zero near the roots (the plants cannot transpire enough water to draw more of those nutrients near the roots). The very steep concentration gradient is of greater influence in the movement of those ions than is the movement of those by mass flow.^[355] The movement by mass flow requires the transpiration of water from the plant causing water and solution ions to also move toward the roots. Movement by root interception is slowest as the plants must extend their roots.

Plants move ions out of their roots in an effort to move nutrients in from the soil. Hydrogen H^+ is exchanged for other cations, and carbonate (HCO_3^-) and hydroxide (OH^-) anions are exchanged for nutrient anions. As plant roots remove nutrients from the soil water solution, they are replenished as other ions move off of clay and humus (by ion exchange or desorption), are added from the weathering of soil minerals, and are released by the decomposition of soil organic matter. Plants derive a large proportion of their anion nutrients from decomposing organic matter, which typically holds about 95 percent of the soil nitrogen, 5 to 60 percent of the soil phosphorus and about 80 percent of the soil sulfur. Where crops are produced, the replenishment of nutrients in the soil must usually be augmented by the addition of fertilizer or organic matter.^[354]

Because nutrient uptake is an active metabolic process, conditions that inhibit root metabolism may also inhibit nutrient uptake. Examples of such conditions include waterlogging or soil compaction resulting in poor soil aeration, excessively high or low soil temperatures, and above-ground conditions that result in low translocation of sugars to plant roots.^[356]

Carbon

Plants obtain their carbon from atmospheric carbon dioxide. About 45% of a plant's dry mass is carbon; plant residues typically have a carbon to nitrogen ratio (C/N) of between 13:1 and 100:1. As the soil organic material is digested by arthropods and micro-organisms, the C/N decreases as the carbonaceous material is metabolized and carbon dioxide (CO₂) is released as a byproduct which then finds its way out of the soil and into the atmosphere. The nitrogen is sequestered in the bodies of the living matter of those decomposing organisms and so it builds up in the soil. Normal CO₂ concentration in the atmosphere is 0.03%, this can be the factor limiting plant growth. In a field of maize on a still day during high light conditions in the growing season, the CO₂ concentration drops very low, but under such conditions the crop could use up to 20 times the normal concentration. The respiration of CO₂ by soil micro-organisms decomposing soil organic matter contributes an important amount of CO₂ to the photosynthesising plants. Within the soil, CO₂ concentration is 10 to 100 times that of atmospheric levels but may rise to toxic levels if the soil porosity is low or if diffusion is impeded by flooding.^{[357][347][358]}

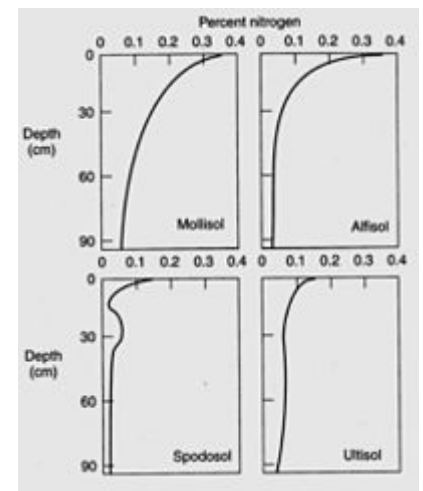


Measuring soil respiration in the field using an SRS2000 system.

Nitrogen

Nitrogen is the most critical element obtained by plants from the soil and nitrogen deficiency often limits plant growth.^[359] Plants can use the nitrogen as either the ammonium cation (NH₄⁺) or the anion nitrate (NO₃⁻). Usually, most of the nitrogen in soil is bound within organic compounds that make up the soil organic matter, and must be mineralized to the ammonium or nitrate form before it can be taken up by most plants. The total nitrogen content depends largely on the soil organic matter content, which in turn depends on the climate, vegetation, topography, age and soil management. Soil nitrogen typically decreases by 0.2 to 0.3% for every temperature increase by 10 °C. Usually, grassland soils contain more soil nitrogen than forest soils. Cultivation decreases soil nitrogen by exposing soil organic matter to decomposition by microorganisms, and soils under no-tillage maintain more soil nitrogen than tilled soils.

Some micro-organisms are able to metabolise organic matter and release ammonium in a process called *mineralisation*. Others take free ammonium and oxidise it to nitrate. Nitrogen-fixing bacteria are capable of metabolising N₂ into the form of ammonia in a process called nitrogen fixation. Both ammonium and nitrate can be *immobilized* by their incorporation into the microbes' living cells, where it is temporarily sequestered in the form of amino acids and protein. Nitrate may also be lost from the soil when bacteria metabolise it to the gases N₂ and N₂O. The loss of gaseous forms of nitrogen to the atmosphere due to microbial action is called *denitrification*. Nitrogen may also be *leached* from the soil if it is in the form of nitrate or lost to the atmosphere as ammonia due to a chemical reaction of ammonium with alkaline soil by way of a process called *volatilisation*. Ammonium may also be sequestered in clay by *fixation*. A small amount of nitrogen is added to soil by rainfall.^{[351][360][361][362]}



Generalization of percent soil nitrogen by soil order

Nitrogen gains

In the process of mineralisation, microbes feed on organic matter, releasing ammonia (NH₃), ammonium (NH₄⁺) and other nutrients. As long as the carbon to nitrogen ratio (C/N) of fresh residues in the soil is above 30:1, nitrogen will be in short supply and other bacteria will feed on the ammonium and incorporate its nitrogen into their cells in the immobilization process. In that form the nitrogen is said to be *immobilised*. Later, when such bacteria die, they too are *mineralised* and some of the nitrogen is released as ammonium and nitrate. If the C/N is less than 15, ammonia is freed to the soil, where it may be used by bacteria which oxidise it to nitrate (nitrification). Bacteria may on average add 25 pounds (11 kg) nitrogen per acre, and in an unfertilised field, this is the most

important source of usable nitrogen. In a soil with 5% organic matter perhaps 2 to 5% of that is released to the soil by such decomposition. It occurs fastest in warm, moist, well aerated soil. The mineralisation of 3% of the organic material of a soil that is 4% organic matter overall, would release 120 pounds (54 kg) of nitrogen as ammonium per acre.^[363]

Carbon/Nitrogen Ratio of Various Organic Materials^[364]

Organic Material	C:N Ratio
Alfalfa	13
Bacteria	4
Clover, green sweet	16
Clover, mature sweet	23
Fungi	9
Forest litter	30
Humus in warm cultivated soils	11
Legume-grass hay	25
Legumes (alfalfa or clover), mature	20
Manure, cow	18
Manure, horse	16–45
Manure, human	10
Oat straw	80
Straw, cornstalks	90
Sawdust	250

In nitrogen fixation, rhizobium bacteria convert N_2 to ammonia (NH_3). Rhizobia share a symbiotic relationship with host plants, since rhizobia supply the host with nitrogen and the host provides rhizobia with nutrients and a safe environment. It is estimated that such symbiotic bacteria in the root nodules of legumes add 45 to 250 pounds of nitrogen per acre per year, which may be sufficient for the crop. Other, free-living nitrogen-fixing bacteria and blue-green algae live independently in the soil and release nitrate when their dead bodies are converted by way of mineralisation.^[365]

Some amount of usable nitrogen is fixed by lightning as nitric oxide (NO) and nitrogen dioxide (NO_2^-). Nitrogen dioxide is soluble in water to form nitric acid (HNO_3) solution of H^+ and NO_3^- . Ammonia, NH_3 , previously released from the soil or from combustion, may fall with precipitation as nitric acid at a rate of about five pounds nitrogen per acre per year.^[366]

Nitrogen sequestration

When bacteria feed on soluble forms of nitrogen (ammonium and nitrate), they temporarily sequester that nitrogen in their bodies in a process called *immobilisation*. At a later time when those bacteria die, their nitrogen may be released as ammonium by the processes of mineralisation.

Protein material is easily broken down, but the rate of its decomposition is slowed by its attachment to the crystalline structure of clay and when trapped between the clay layers. The layers are small enough that bacteria cannot enter. Some organisms can exude extracellular enzymes that can act on the sequestered proteins. However, those enzymes too may be trapped on the clay crystals.

Ammonium fixation occurs when ammonium pushes potassium ions from between the layers of clay such as illite or montmorillonite. Only a small fraction of soil nitrogen is held this way.^[324]

Nitrogen losses

Usable nitrogen may be lost from soils when it is in the form of nitrate, as it is easily leached. Further losses of nitrogen occur by denitrification, the process whereby soil bacteria convert nitrate (NO_3^-) to nitrogen gas, N_2 or N_2O . This occurs when poor soil aeration limits free oxygen, forcing bacteria to use the oxygen in nitrate for their respiratory process. Denitrification increases when oxidisable organic material is available and when soils are warm and slightly acidic. Denitrification may vary throughout a soil as the aeration varies from place to place. Denitrification may cause the loss of 10 to 20 percent of the available nitrates within a day and when conditions are favourable to that process, losses of up to 60 percent of nitrate applied as fertiliser may occur^[367]

Ammonium volatilisation occurs when ammonium reacts chemically with an alkaline soil, converting NH_4^+ to NH_3 . The application of ammonium fertiliser to such a field can result in volatilisation losses of as much as 30 percent.^[368]

Phosphorus

After nitrogen, phosphorus is probably the element most likely to be deficient in soils. The soil mineral apatite is the most common mineral source of phosphorus. While there is on average 1000 lb of phosphorus per acre in the soil, it is generally unavailable in the form of phosphates of low solubility. Total phosphorus is about 0.1 percent by weight of the soil, but only one percent of that is available. Of the part available, more than half comes from the mineralisation of organic matter. Agricultural fields may need to be fertilised to make up for the phosphorus that has been removed in the crop.^[330]

When phosphorus does form solubilised ions of H_2PO_4^- , they rapidly form insoluble phosphates of calcium or hydrous oxides of iron and aluminum. Phosphorus is largely immobile in the soil and is not leached but actually builds up in the surface layer if not cropped. The application of soluble fertilisers to soils may result in zinc deficiencies as zinc phosphates form. Conversely, the application of zinc to soils may immobilise phosphorus again as zinc phosphate. Lack of phosphorus may interfere with the normal opening of the plant leaf stomata, resulting in plant temperatures 10 percent higher than normal. Phosphorus is most available when soil pH is 6.5 in mineral soils and 5.5 in organic soils.^[368]

Potassium

The amount of potassium in a soil may be as much as 80,000 lb per acre-foot, of which only 150 lb is available for plant growth. Common mineral sources of potassium are the mica biotite and potassium feldspar, KAlSi_3O_8 . When solubilised, half will be held as exchangeable cations on clay while the other half is in the soil water solution. Potassium fixation often occurs when soils dry and the potassium is bonded between layers of illite clay. Under certain conditions, dependent on the soil texture, intensity of drying, and initial amount of exchangeable potassium, the fixed percentage may be as much as 90 percent within ten minutes. Potassium may be leached from soils low in clay^{[369][370]}

Calcium

Calcium is one percent by weight of soils and is generally available but may be low as it is soluble and can be leached. It is thus low in sandy and heavily leached soil or strongly acidic mineral soil. Calcium is supplied to the plant in the form of exchangeable ions and moderately soluble minerals. Calcium is more available on the soil colloids than is potassium because the common mineral calcite, CaCO_3 , is more soluble than potassium-bearing minerals.^[371]

Magnesium

Magnesium is one of the dominant exchangeable cations in most soils (as are calcium and potassium). Primary minerals that weather to release magnesium include hornblende, biotite and vermiculite. Soil magnesium concentrations are generally sufficient for optimal plant growth, but highly weathered and sandy soils may be magnesium deficient due to leaching by heavy precipitation.^{[351][372]}

Sulfur

Most sulfur is made available to plants, like phosphorus, by its release from decomposing organic matter.^[372] Deficiencies may exist in some soils (especially sandy soils) and if cropped, sulfur needs to be added. The application of large quantities of nitrogen to fields that have marginal amounts of sulfur may cause sulfur deficiency in the rapidly growing plants by the plant's growth outpacing the supply of sulfur. A 15-ton crop of onions uses up to 19 lb of sulfur and 4 tons of alfalfa uses 15 lb per acre. Sulfur abundance varies with depth. In a sample of soils in Ohio, United States, the sulfur abundance varied with depths, 0-6 inches, 6-12 inches, 12-18 inches, 18-24 inches in the amounts: 1056, 830, 686, 528 lb per acre respectively.^[373]

Micronutrients

The micronutrients essential for plant life, in their order of importance, include iron,^[374] manganese,^[375] zinc,^[376] copper,^[377] boron,^[378] chlorine^[379] and molybdenum.^[380] The term refers to plants' needs, not to their abundance in soil. They are required in very small amounts but are essential to plant health in that most are required parts of some enzyme system which speeds up plants' metabolisms. They are generally available in the mineral component of the soil, but the heavy application of phosphates can cause a deficiency in zinc and iron by the formation of insoluble zinc and iron phosphates. Iron deficiency may also result from excessive amounts of heavy metals or calcium minerals (lime) in the soil. Excess amounts of soluble boron, molybdenum and chloride are toxic.^{[381][382]}

Non-essential nutrients

Nutrients which enhance the health but whose deficiency does not stop the life cycle of plants include: cobalt, strontium, vanadium, silicon and nickel.^[383] As their importance are evaluated they may be added to the list of essential plant nutrients.

Soil organic matter

Soil organic matter is made up of organic compounds and includes plant, animal and microbial material, both living and dead. A typical soil has a biomass composition of 70% microorganisms, 22% macrofauna, and 8% roots. The living component of an acre of soil may include 900 lb of earthworms, 2400 lb of fungi, 1500 lb of bacteria, 133 lb of protozoa and 890 lb of arthropods and algae.^[384]

A small part of the organic matter consists of the living cells such as bacteria, molds, and actinomycetes that work to break down the dead organic matter. Were it not for the action of these micro-organisms, the entire carbon dioxide part of the atmosphere would be sequestered as organic matter in the soil.

Chemically, organic matter is classed as follows:

1. Polysaccharides
 1. cellulose
 2. hemicellulose
 3. starch
 4. pectin
2. Lignins
3. Proteins

Most living things in soils, including plants, insects, bacteria, and fungi, are dependent on organic matter for nutrients and/or energy. Soils have organic compounds in varying degrees of decomposition which rate is dependent on the temperature, soil moisture, and aeration. Bacteria and fungi feed on the raw organic matter, which are fed upon by amoebas, which in turn are fed upon by nematodes and arthropods. Organic matter holds soils open, allowing the infiltration of air and water, and may hold as much as twice its weight in water. Many soils, including desert and rocky-gravel soils, have little or no organic matter. Soils that are all organic matter, such as peat (histosols), are infertile.^[385] In its earliest stage of decomposition, the original organic material is often called raw organic matter. The final stage of decomposition is called humus.

In grassland, much of the organic matter added to the soil is from the deep, fibrous, grass root systems. By contrast, tree leaves falling on the forest floor are the principal source of soil organic matter in the forest. Another difference is the frequent occurrence in the grasslands of fires that destroy large amounts of aboveground material but stimulate even greater contributions from roots. Also, the much greater acidity under any forests inhibits the action of certain soil organisms that otherwise would mix much of the surface litter into the mineral soil. As a result, the soils under grasslands generally develop a thicker A horizon with a deeper distribution of organic matter than in comparable soils under forests, which characteristically store most of their organic matter in the forest floor (O horizon) and thin A horizon.

Humus

Humus refers to organic matter that has been decomposed by soil flora and fauna to the point where it is resistant to further breakdown. Humus usually constitutes only five percent of the soil or less by volume, but it is an essential source of nutrients and adds important textural qualities crucial to soil health and plant growth. Humus also holds bits of undecomposed organic matter which feed arthropods and worms which further improve the soil. The end product, humus, is soluble in water and forms a weak acid that can attack silicate minerals.^[386] Humus is a colloid with a high cation and anion exchange capacity that on a dry weight basis is many times greater than that of clay colloids. It also acts as a buffer, like clay, against changes in pH and soil moisture.

Humic acids and fulvic acids, which begin as raw organic matter, are important constituents of humus. After the death of plants and animals, microbes begin to feed on the residues, resulting finally in the formation of humus. With decomposition, there is a reduction of water-soluble constituents, cellulose and hemicellulose, and nutrients such as nitrogen, phosphorus, and sulfur. As the residues break down, only stable molecules made of aromatic carbon rings, oxygen and hydrogen remain in the form of humins, lignin and lignin complexes collectively called humus. While the structure of humus has few nutrients, it is able to attract and hold cation and anion nutrients by weak bonds that can be released into the soil solution in response to changes in soil pH.

Lignin is resistant to breakdown and accumulates within the soil. It also reacts with amino acids, which further increases its resistance to decomposition, including enzymatic decomposition by microbes. Fats and waxes from plant matter have some resistance to decomposition and persist in soils for a while. Clay soils often have higher organic contents that persist longer than soils without clay as the organic molecules adhere to and are stabilised by the clay. Proteins normally decompose readily, but when bound to clay particles, they become more resistant to decomposition. Clay particles also absorb the enzymes exuded by microbes which would normally break down proteins. The addition of organic matter to clay soils can render that organic matter and any added nutrients inaccessible to plants and microbes for many years. High soil tannin (polyphenol) content can cause nitrogen to be sequestered in proteins or cause nitrogen immobilisation.^{[387][388]}

Humus formation is a process dependent on the amount of plant material added each year and the type of base soil. Both are affected by climate and the type of organisms present. Soils with humus can vary in nitrogen content but typically have 3 to 6 percent nitrogen. Raw organic matter, as a reserve of nitrogen and phosphorus, is a vital component affecting soil fertility.^[385] Humus also absorbs water, and expands and shrinks between dry and wet states, increasing soil porosity. Humus is less stable than the soil's mineral constituents, as it is reduced by microbial decomposition, and over time its concentration diminishes without the addition of new organic matter. However, humus may persist over centuries if not millennia.

Climate and organic matter

The production, accumulation and degradation of organic matter are greatly dependent on climate. Temperature, soil moisture and topography are the major factors affecting the accumulation of organic matter in soils. Organic matter tends to accumulate under wet or cold conditions where decomposer activity is impeded by low temperature^[389] or excess moisture which results in anaerobic conditions.^[390] Conversely, excessive rain and high temperatures of tropical climates enables rapid decomposition of organic matter and leaching of plant nutrients; forest ecosystems on these soils rely on efficient recycling of nutrients and plant matter to maintain their productivity.^[391] Excessive slope may encourage the erosion of the top layer of soil which holds most of the raw organic material that would otherwise eventually become humus.

Plant residue in soil

Cellulose and hemicellulose undergo fast decomposition by fungi and bacteria, with a half-life of 12–18 days in a temperate climate.^[392] Brown rot fungi can decompose the cellulose and hemicellulose, leaving the lignin and phenolic compounds behind. Starch, which is an energy storage system for plants, undergoes fast decomposition by bacteria and fungi. Lignin consists of polymers composed of 500 to 600 units with a highly branched, amorphous structure. Lignin undergoes very slow decomposition, mainly by white rot fungi and actinomycetes; its half-life under temperate conditions is about six months.^[392]

Soil horizons

A horizontal layer of the soil, whose physical features, composition and age are distinct from those above and beneath, is referred to as a soil horizon. The naming of a horizon is based on the type of material of which it is composed. Those materials reflect the duration of specific processes of soil formation. They are labelled using a shorthand notation of letters and numbers^[393] which describe the horizon in terms of its colour, size, texture, structure, consistency, root quantity, pH, voids, boundary characteristics and presence of nodules or concretions.^[394] No soil profile has all the major horizons. Some may have only one horizon.

The exposure of parent material to favourable conditions produces mineral soils that are marginally suitable for plant growth. That growth often results in the accumulation of organic residues. The accumulated organic layer called the O horizon produces a more active soil due to the effect of the organisms that live within it. Organisms colonise and break down organic materials, making available nutrients upon which other plants and animals can live. After sufficient time, humus moves downward and is deposited in a distinctive organic surface layer called the A horizon.

Classification

Soil is classified into categories in order to understand relationships between different soils and to determine the suitability of a soil for a particular use. One of the first classification systems was developed by Russian scientist Dokuchaev around 1880. It was modified a number of times by American and European researchers, and developed into the system commonly used until the 1960s. It was based on the idea that soils have a particular morphology based on the materials and factors that form them. In the 1960s, a different classification system began to emerge which focused on soil morphology instead of parental materials and soil-forming factors. Since then it has undergone further modifications. The World Reference Base for Soil Resources (WRB)^[395] aims to establish an international reference base for soil classification.

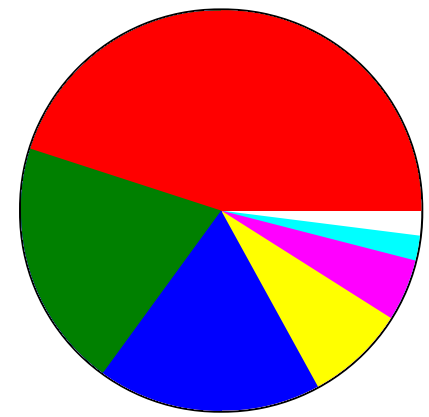
Soil classification systems

Australia

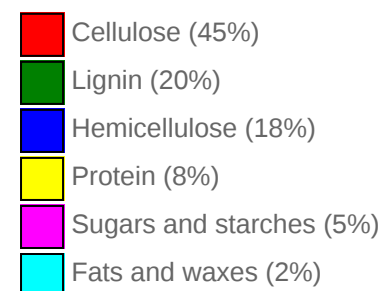
There are fourteen soil orders at the top level of the Australian Soil Classification. They are: Anthrosols, Organosols, Podosols, Vertosols, Hydrosols, Kurosols, Sodosols, Chromsols, Calcarosols, Ferrosols, Dermosols, Kandosols, Rudosols and Entosols.

European Union

The EU's soil taxonomy is based on a new standard soil classification in the World Reference Base for Soil Resources produced by the UN's Food and Agriculture Organization.^[396] According to this, the major soils in the European Union are:



Typical types and percentages of plant residue components



- Acrisols
- Albeluvisols
- Andosols
- Anthrosols
- Arenosols
- Calcisols
- Cambisols
- Chernozems
- Fluvisols
- Gleysols
- Gypsisols
- Histosols
- Kastanozems
- Leptosols
- Luvisols
- Phaeozems
- Planosols
- Podzols
- Regosols
- Solonchaks
- Solonetzes
- Umbrisols
- Vertisols^[396]

United States

A taxonomy is an arrangement in a systematic manner; the USDA soil taxonomy has six levels of classification. They are, from most general to specific: order, suborder, great group, subgroup, family and series. Soil properties that can be measured quantitatively are used in this classification system – they include: depth, moisture, temperature, texture, structure, cation exchange capacity, base saturation, clay mineralogy, organic matter content and salt content. There are 12 soil orders (the top hierarchical level) in soil taxonomy.^{[397][398]} The names of the orders end with the suffix *-sol*. The criteria for the different soil orders include properties that reflect major differences in the genesis of soils.^[399] The orders are:

- Alfisol – soils with aluminium and iron. They have horizons of clay accumulation, and form where there is enough moisture and warmth for at least three months of plant growth. They constitute 10% of soils worldwide.
- Andisol – volcanic ash soils. They are young and very fertile. They cover 1% of the world's ice-free surface.
- Aridisol – dry soils forming under desert conditions which have fewer than 90 consecutive days of moisture during the growing season and are nonleached. They include nearly 12% of soils on Earth. Soil formation is slow and accumulated organic matter is scarce. They may have subsurface zones of caliche or duripan. Many aridisols have well-developed Bt horizons showing clay movement from past periods of greater moisture.
- Entisol – recently formed soils that lack well-developed horizons. Commonly found on unconsolidated river and beach sediments of sand and clay or volcanic ash, some have an A horizon on top of bedrock. They are 18% of soil worldwide.
- Gelisol – permafrost soils with permafrost within two metres of the surface or gelic materials and permafrost within one metre. They constitute 9% of soils worldwide.
- Histosol – organic soils, formerly called bog soils, are 1% of soils worldwide.
- Inceptisol – young soils. They have subsurface horizon formation but show little eluviation and illuviation. They constitute 15% of soils worldwide.
- Mollisol – soft, deep, dark fertile soil formed in grasslands and some hardwood forests with very thick A horizons. They are 7% of soils worldwide.
- Oxisol – are heavily weathered, are rich in iron and aluminum oxides (sesquioxides) or kaolin but low in silica. They have only trace nutrients due to heavy tropical rainfall and high temperatures and low CEC of the remaining clays. They are 8% of soils worldwide.
- Spodosol – acid soils with organic colloid layer complexed with iron and aluminium leached from a layer above. They are typical soils of coniferous and deciduous forests in cooler climates. They constitute 4% of soils worldwide.
- Ultisol – acid soils in the humid tropics and subtropics, which are depleted in calcium, magnesium and potassium (important plant nutrients). They are highly weathered, but not as weathered as Oxisols. They make up 8% of the

soil worldwide.

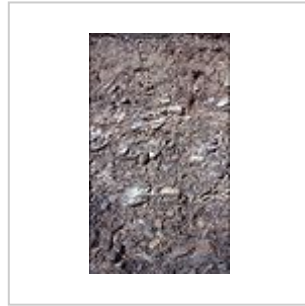
- **Vertisol** – inverted soils. They are clay-rich and tend to swell when wet and shrink upon drying, often forming deep cracks into which surface layers can fall. They are difficult to farm or to construct roads and buildings due to their high expansion rate. They constitute 2% of soils worldwide.



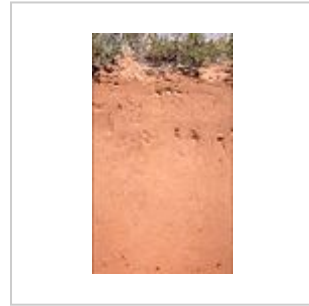
Alfisol



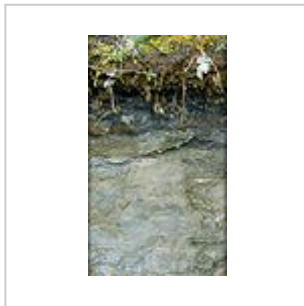
Andisol



Aridisol



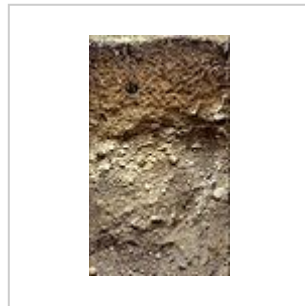
Entisol



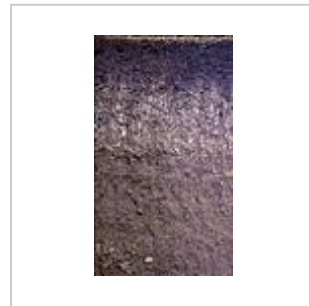
Gelisol



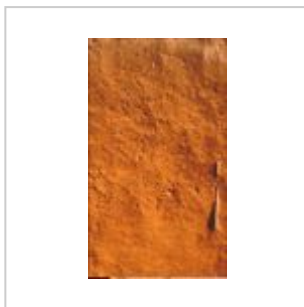
Histisol



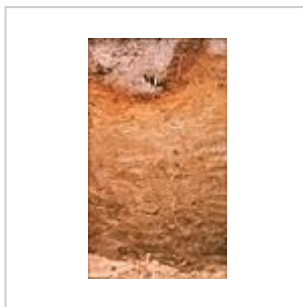
Inceptisol



Mollisol



Oxisol



Spodosol



Ultisol



Vertisol

The percentages listed above^[400] are for land area free of ice. "Soils of Mountains", which constitute the balance (11.6%), have a mixture of those listed above, or are classified as "Rugged Mountains" which have no soil.

The above soil orders in sequence of increasing degree of development are Entisols, Inceptisols, Aridisols, Mollisols, Alfisols, Spodosols, Ultisols, and Oxisols. Histosols and Vertisols may appear in any of the above at any time during their development.

The **soil suborders** within an order are differentiated on the basis of soil properties and horizons which depend on soil moisture and temperature. Forty-seven suborders are recognized in the United States.^[401]

The **soil great group** category is a subdivision of a suborder in which the kind and sequence of soil horizons distinguish one soil from another. About 185 great groups are recognized in the United States. Horizons marked by clay, iron, humus and hard pans and soil features such as the expansion-contraction of clays (that produce self-mixing provided by clay), temperature, and marked quantities of various salts are used as distinguishing features.^[401]

The great group categories are divided into three kinds of **soil subgroups**: typic, intergrade and extragrade. A typic subgroup represents the basic or 'typical' concept of the great group to which the described subgroup belongs. An **intergrade** subgroup describes the properties that suggest how it grades towards (is similar to) soils of other soil great groups, suborders or orders. These properties

are not developed or expressed well enough to cause the soil to be included within the great group towards which they grade, but suggest similarities. Extragrade features are aberrant properties which prevent that soil from being included in another soil classification. About 1,000 soil subgroups are defined in the United States.^[401]

A **soil family** category is a group of soils within a subgroup and describes the physical and chemical properties which affect the response of soil to agricultural management and engineering applications. The principal characteristics used to differentiate soil families include texture, mineralogy, pH, permeability, structure, consistency, the locale's precipitation pattern, and soil temperature. For some soils the criteria also specify the percentage of silt, sand and coarse fragments such as gravel, cobbles and rocks. About 4,500 soil families are recognised in the United States.^[402]

A family may contain several **soil series** which describe the physical location using the name of a prominent physical feature such as a river or town near where the soil sample was taken. An example would be Merrimac for the Merrimack River in New Hampshire. More than 14,000 soil series are recognised in the United States. This permits very specific descriptions of soils.^[403]

A **soil phase of series**, originally called 'soil type' describes the soil surface texture, slope, stoniness, saltiness, erosion, and other conditions.^[403]

Uses

Soil is used in agriculture, where it serves as the anchor and primary nutrient base for plants; however, as demonstrated by hydroponics, it is not essential to plant growth if the soil-contained nutrients can be dissolved in a solution. The types of soil and available moisture determine the species of plants that can be cultivated.

Soil material is also a critical component in the mining, construction and landscape development industries.^[404] Soil serves as a foundation for most construction projects. The movement of massive volumes of soil can be involved in surface mining, road building and dam construction. Earth sheltering is the architectural practice of using soil for external thermal mass against building walls. Many building materials are soil based.

Soil resources are critical to the environment, as well as to food and fibre production. Soil provides minerals and water to plants. Soil absorbs rainwater and releases it later, thus preventing floods and drought. Soil cleans water as it percolates through it. Soil is the habitat for many organisms: the major part of known and unknown biodiversity is in the soil, in the form of invertebrates (earthworms, woodlice, millipedes, centipedes, snails, slugs, mites, springtails, enchytraeids, nematodes, protists), bacteria, archaea, fungi and algae; and most organisms living above ground have part of them (plants) or spend part of their life cycle (insects) below-ground. Above-ground and below-ground biodiversities are tightly interconnected,^{[405][406]} making soil protection of paramount importance for any restoration or conservation plan.

The biological component of soil is an extremely important carbon sink since about 57% of the biotic content is carbon. Even on desert crusts, cyanobacteria, lichens and mosses capture and sequester a significant amount of carbon by photosynthesis. Poor farming and grazing methods have degraded soils and released much of this sequestered carbon to the atmosphere. Restoring the world's soils could offset some of the huge increase in greenhouse gases causing global warming, while improving crop yields and reducing water needs.^{[407][408][409]}

Waste management often has a soil component. Septic drain fields treat septic tank effluent using aerobic soil processes. Landfills use soil for daily cover. Land application of waste water relies on soil biology to aerobically treat BOD.

Organic soils, especially peat, serve as a significant fuel resource; but wide areas of peat production, such as sphagnum bogs, are now protected because of patrimonial interest.

Geophagy is the practice of eating soil-like substances. Both animals and human cultures occasionally consume soil for medicinal, recreational, or religious purposes. It has been shown that some monkeys consume soil, together with their preferred food (tree foliage and fruits), in order to alleviate tannin toxicity.^[410]

Soils filter and purify water and affect its chemistry. Rain water and pooled water from ponds, lakes and rivers percolate through the soil horizons and the upper rock strata, thus becoming groundwater. Pests (viruses) and pollutants, such as persistent organic pollutants (chlorinated pesticides, polychlorinated biphenyls), oils (hydrocarbons), heavy metals (lead, zinc, cadmium), and excess nutrients (nitrates, sulfates, phosphates) are filtered out by the soil.^[411] Soil organisms metabolise them or immobilise them in their biomass and necromass,^[412] thereby incorporating them into stable humus.^[413] The physical integrity of soil is also a prerequisite for avoiding landslides in rugged landscapes.^[414]

Degradation

Land degradation^[415] refers to a human-induced or natural process which impairs the capacity of land to function. Soils degradation involves the acidification, contamination, desertification, erosion or salination.

Soil acidification is beneficial in the case of alkaline soils, but it degrades land when it lowers crop productivity and increases soil vulnerability to contamination and erosion. Soils are often initially acid because their parent materials were acid and initially low in the basic cations (calcium, magnesium, potassium and sodium). Acidification occurs when these elements are leached from the soil profile by rainfall or by the harvesting of forest or agricultural crops. Soil acidification is accelerated by the use of acid-forming nitrogenous fertilizers and by the effects of acid precipitation.

Soil contamination at low levels is often within a soil's capacity to treat and assimilate waste material. Soil biota can treat waste by transforming it; soil colloids can adsorb the waste material. Many waste treatment processes rely on this treatment capacity. Exceeding treatment capacity can damage soil biota and limit soil function. Derelict soils occur where industrial contamination or other development activity damages the soil to such a degree that the land cannot be used safely or productively. Remediation of derelict soil uses principles of geology, physics, chemistry and biology to degrade, attenuate, isolate or remove soil contaminants to restore soil functions and values. Techniques include leaching, air sparging, chemical amendments, phytoremediation, bioremediation and natural degradation.

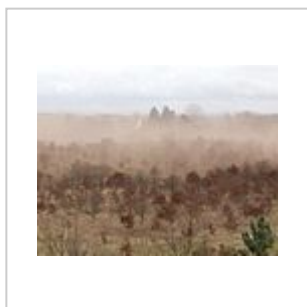
Desertification is an environmental process of ecosystem degradation in arid and semi-arid regions, often caused by human activity. It is a common misconception that droughts cause desertification. Droughts are common in arid and semiarid lands. Well-managed lands can recover from drought when the rains return. Soil management tools include maintaining soil nutrient and organic matter levels, reduced tillage and increased cover. These practices help to control erosion and maintain productivity during periods when moisture is available. Continued land abuse during droughts, however, increases land degradation. Increased population and livestock pressure on marginal lands accelerates desertification.

Erosion of soil is caused by water, wind, ice, and movement in response to gravity. More than one kind of erosion can occur simultaneously. Erosion is distinguished from weathering, since erosion also transports eroded soil away from its place of origin (soil in transit may be described as sediment). Erosion is an intrinsic natural process, but in many places it is greatly increased by human activity, especially poor land use practices. These include agricultural activities which leave the soil bare during times of heavy rain or strong winds, overgrazing, deforestation, and improper construction activity. Improved management can limit erosion. Soil conservation techniques which are employed include changes of land use (such as replacing erosion-prone crops with grass or other soil-binding plants), changes to the timing or type of agricultural operations, terrace building, use of erosion-suppressing cover materials (including cover crops and other plants), limiting disturbance during construction, and avoiding construction during erosion-prone periods.

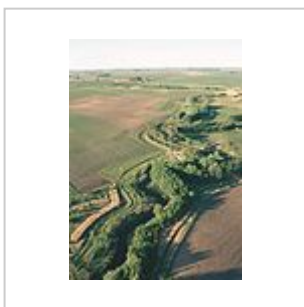
A serious and long-running water erosion problem occurs in China, on the middle reaches of the Yellow River and the upper reaches of the Yangtze River. From the Yellow River, over 1.6 billion tons of sediment flow each year into the ocean. The sediment originates primarily from water erosion (gully erosion) in the Loess Plateau region of northwest China.

Soil piping is a particular form of soil erosion that occurs below the soil surface. It causes levee and dam failure, as well as sink hole formation. Turbulent flow removes soil starting at the mouth of the seep flow and the subsoil erosion advances up-gradient.^[416] The term sand boil is used to describe the appearance of the discharging end of an active soil pipe.^[417]

Soil salination is the accumulation of free salts to such an extent that it leads to degradation of the agricultural value of soils and vegetation. Consequences include corrosion damage, reduced plant growth, erosion due to loss of plant cover and soil structure, and water quality problems due to sedimentation. Salination occurs due to a combination of natural and human-caused processes. Arid conditions favour salt accumulation. This is especially apparent when soil parent material is saline. Irrigation of arid lands is especially problematic.^[418] All irrigation water has some level of salinity. Irrigation, especially when it involves leakage from canals and overirrigation in the field, often raises the underlying water table. Rapid salination occurs when the land surface is within the capillary fringe of saline groundwater. Soil salinity control involves watertable control and flushing with higher levels of applied water in combination with tile drainage or another form of subsurface drainage.^{[419][420]}



Desertification



Erosion control

Reclamation

Soils which contain high levels of particular clays, such as smectites, are often very fertile. For example, the smectite-rich clays of Thailand's Central Plains are among the most productive in the world.

Many farmers in tropical areas, however, struggle to retain organic matter in the soils they work. In recent years, for example, productivity has declined in the low-clay soils of northern Thailand. Farmers initially responded by adding organic matter from termite mounds, but this was unsustainable in the long-term. Scientists experimented with adding bentonite, one of the smectite family of clays, to the soil. In field trials, conducted by scientists from the International Water Management Institute in cooperation with Khon Kaen University and local farmers, this had the effect of helping retain water and nutrients. Supplementing the farmer's usual practice with a single application of 200 kg bentonite per rai (6.26 rai = 1 hectare) resulted in an average yield increase of 73%. More work showed that applying bentonite to degraded sandy soils reduced the risk of crop failure during drought years.

In 2008, three years after the initial trials, IWMI scientists conducted a survey among 250 farmers in northeast Thailand, half of whom had applied bentonite to their fields. The average improvement for those using the clay addition was 18% higher than for non-clay users. Using the clay had enabled some farmers to switch to growing vegetables, which need more fertile soil. This helped to increase their income. The researchers estimated that 200 farmers in northeast Thailand and 400 in Cambodia had adopted the use of clays, and that a further 20,000 farmers were introduced to the new technique.^[421]

If the soil is too high in clay, adding gypsum, washed river sand and organic matter will balance the composition. Adding organic matter (like ramial chipped wood for instance) to soil which is depleted in nutrients and too high in sand will boost its quality.^[422]

See also

- Acid sulfate soil
- Agrophysics
- Alkaline soil
- Biochar
- Bulk soil
- Geoponic
- Mycorrhizal fungi and soil carbon storage
- Nitrogen cycle
- Red Mediterranean soil
- Saline soil
- Shrink-swell capacity
- Soil management

- [Factors affecting permeability of soils](#)
- [Index of soil-related articles](#)
- [Laterite](#)
- [Manure](#)
- [Mineral matter in plants](#)
- [Soil zoology](#)
- [World Soil Museum](#)

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- [LandIS Soils Data for England and Wales](#) a pay source for GIS data on the soils of England and Wales and soils data source; they charge a handling fee to researchers.

External links

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|---|--|
| ▪ The Soil Water Compendium (soil water content sensors explained) | ▪ Canadian Society of Soil Science |
| ▪ Global Soil Partnership | ▪ Soil Science Society of America |
| ▪ FAO Soils Portal | ▪ USDA-NRCS Web Soil Survey |
| ▪ World Reference Base for Soil Resources | ▪ European Soil Portal (wiki) |
| ▪ ISRIC – World Soil Information (ICSU World Data Centre for Soils) | ▪ National Soil Resources Institute UK |
| ▪ World Soil Library and Maps | ▪ Plant and Soil Sciences eLibrary |
| ▪ Wossac the world soil survey archive and catalogue | ▪ [1] Copies of the reference 'Soil: The Yearbook of Agriculture 1957' in multiple formats |
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