

## Chapter 2

# The Soil Habitat

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## I INTRODUCTION

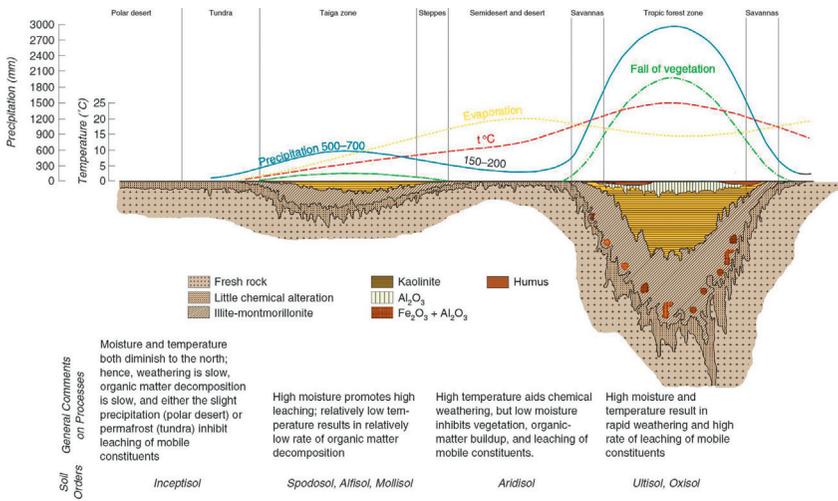
Soil, the naturally occurring unconsolidated mineral and organic material at the earth's surface, provides an essential natural resource for living organisms. It is a central component of the earth's critical zone and deserves special status due to its role in regulating the earth's environment, thus affecting the sustainability of life on the planet. The soil environment is the most complex habitat on earth. This complexity governs soil biodiversity, as soil is estimated to contain one-third of all living organisms and regulates the activity of the organisms responsible for ecosystem functioning and evolution. The concept that the earth's physicochemical properties are tightly coupled to the activity of the living organisms it supports was proposed in the early 1970s by James Lovelock and Lynn Margulis as the Gaia hypothesis. They theorized that the earth behaves as a superorganism, with an intrinsic ability to control its own climate and chemistry and thus maintain an environment favorable for life. Soil has the intrinsic ability to both support terrestrial life and provide a habitat for the interdependent existence and evolution of organisms living within it. In 2011,

scientists embarked on a Global Soil Biodiversity Initiative to assess soil life in all biomes across the globe for the essential ecosystem services that soils provide (i.e., plant biomass production, decomposition, and nutrient cycling) and to identify where soil quality is endangered due to human activities. The ultimate goal of the initiative was to guide environmental policy for sustainable land management (Soil Stories, The Whole Story, <http://youtube.com/watch?v=Ego6LI-IjbY>; see online supplemental material at: <http://booksite.elsevier.com/9780124159556>).

Soils (pedosphere) develop at the interface where organisms (biosphere) interact with rocks and minerals (lithosphere), water (hydrosphere), and air (atmosphere), with climate regulating the intensity of these interactions. In terrestrial ecosystems, the soil affects energy, water and nutrient storage and exchange, and ecosystem productivity. Scientists study soil because of the fundamental need to understand the dynamics of geochemical–biochemical–biophysical interactions at the earth’s surface, especially in light of recent and ongoing changes in global climate and the impact of human activity. Geochemical fluxes between the hydrosphere, atmosphere, and lithosphere take place over the time span of hundreds to millions of years. Within the pedosphere, biologically induced fluxes between the lithosphere, atmosphere, and biosphere take place over a much shorter time frame, hours and days to months, which complicates the study of soils.

The soil habitat is defined as the totality of living organisms inhabiting soil, including plants, animals, and microorganisms, and their abiotic environment. The exact nature of the habitat in which the community of organisms lives is determined by a complex interplay of geology, climate, and plant vegetation. The interactions of rock and parent material with temperature, rainfall, elevation, latitude, exposure to sun and wind, and numerous other factors, over broad geographical regions, environmental conditions, and plant communities, have evolved into the current terrestrial biomes with their associated soils (Fig. 2.1).

Because soils provide such a tremendous range of habitats, they support an enormous biomass, with an estimated  $2.6 \times 10^{29}$  for prokaryotic cells alone, and harbor much of the earth’s genetic diversity. A single gram of soil contains kilometers of fungal hyphae, more than  $10^9$  bacterial and archaeal cells and organisms belonging to tens of thousands of different species. Zones of good aeration may be only millimeters away from areas that are poorly aerated. Areas near the soil surface may be enriched with decaying organic matter and other accessible nutrients, whereas the subsoil may be nutrient poor. The variance of temperature and water content of surface soils is much greater than that of subsoils. The soil solution in some pores may be acidic, yet in others more basic, or may vary in salinity depending on soil mineralogy, location within the landscape, and biological activity. The microenvironment of the surfaces of soil particles, where nutrients are concentrated, is very different from that of the soil solution.



**FIG. 2.1** Environmental factors affecting the distribution of terrestrial biomes and formation of soils along a transect from the equator to the North Polar region (from *Birkeland, 1999*).

## II SOIL GENESIS AND FORMATION OF THE SOIL HABITAT

Soils derived from weathered rocks and minerals are referred to as mineral soils. When plant residues are submerged in water for prolonged periods, biological decay is slowed. Accumulations of organic matter at various stages of decomposition become organic soils and include peat land, muck land, or bogs and fens. Soil can also be formed in coastal tidal marshes or inland water areas supporting plant growth where areas are periodically submerged.

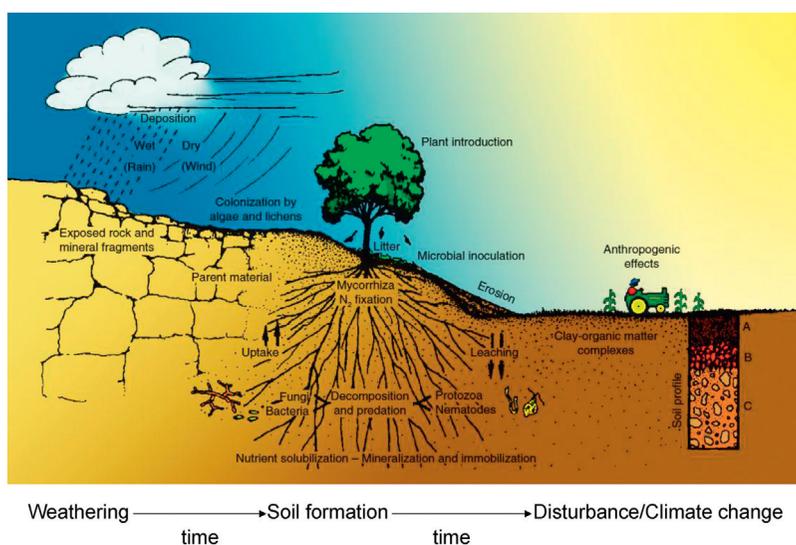
Mineral soils are formed by the physical and chemical weathering of the rocks and minerals brought to the earth's surface by geological processes. They extend from the earth's terrestrial surface into the underlying, relatively unweathered, parent material. The parent material of mineral soils can be the residual material weathered from solid rock masses or the loose, unconsolidated materials that often have been transported from one location and deposited at another by such processes as sedimentation, erosion, and glaciation. The disintegration of rocks into smaller mineral particles is a physical-chemical process brought about by cycles of heating and cooling, freezing and thawing, and also by abrasion from wind, water, and ice masses. Chemical and biochemical weathering processes are enhanced by the presence of water, oxygen, and the organic compounds resulting from biological activity. These reactions convert primary minerals, such as feldspars and micas, to secondary minerals, such as silicate clays and oxides of aluminum, iron, and silica. Soluble constituent elements in inorganic forms provide nutrients to support the growth of various organisms and plants.

The physical and chemical weathering of rocks into fine particles with large surface areas, along with the accompanying release of plant nutrients, initiate the soil-forming process (Fig. 2.2) by providing a habitat for living organisms. The initial colonizers of soil parent material are usually organisms capable of both photosynthesis and  $N_2$  fixation. Intimate root-bacterial/fungal/actinomycetal associations with early plants assist with supplying nutrients and water. Products of the biological decay of organic residues accumulate in the surface soil, which then forms soil organic matter (SOM).

Soil organisms, together with plants, constitute one of the five interactive factors responsible for soil formation. By 1880, Russian and Danish soil scientists had developed the concept of soils as independent natural bodies, each possessing unique properties resulting from parent material, climate, topography, and living matter, interacting over time. The approach for describing soil genesis in the landscape and as a unique biochemical product of organisms participating in the genesis of their own habitat was quantified by Hans Jenny in 1941 in his classic equation of soil forming factors:

$$\text{Soil} = f[\text{Parent material, climate, living organisms, topography, time}] \quad (2.1)$$

It took 10,000-30,000 years to form soils in the glaciated areas; alternatively, soil formation can take hundreds of thousands of years, depending on deposition and erosion processes. (The Five Factors of Soil Formation: <http://www.youtube.com/watch?v=bTzslvAD1Es>; see online supplemental material at: <http://booksite.elsevier.com/9780124159556>).



**FIG. 2.2** Soils are formed as a function of the five interacting factors of climate, parent material, topography, organisms, and time. Human activities have altered soil formation and promoted soil degradation through cultivation and cropping (from Paul and Clark, 1996).

Humans have had a negative effect on soil formation due to some of the agricultural processes they use. The clearing of native vegetation, tillage/cultivation of the surface soil, and cropping for agriculture are known to degrade soils due to promoting erosion and enhancing losses of SOM. Yet alternatively, humans have also contributed to the improvement of soil conditions by installing drainage and irrigation systems and adding amendments of nutrients and organic matter for remediation of mine sites and other areas of exposed parent material. Hans Jenny Memorial Lecture in Soil Science - The Genius of Soil. (<http://www.youtube.com/watch?v=y3q0mg54Li4>; see online supplemental material at: <http://booksite.elsevier.com/9780124159556>).

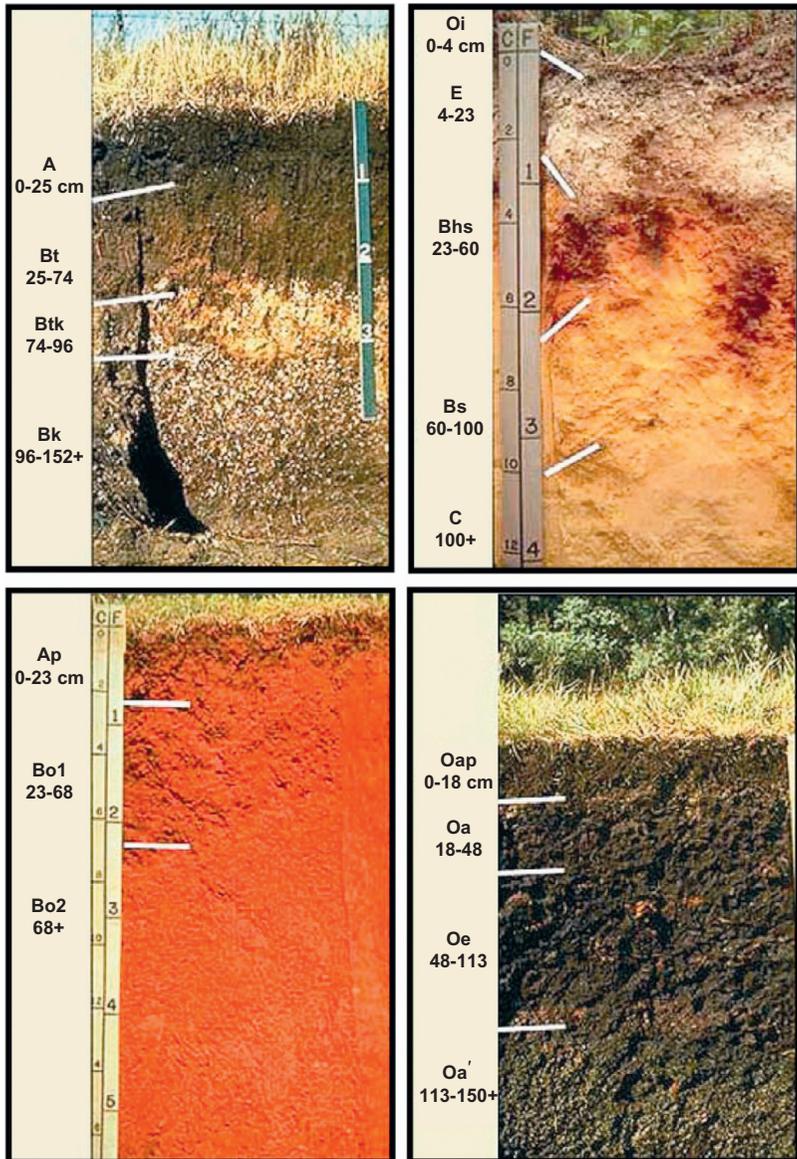
## **A Soil Profile**

During formation, soils develop horizontal layers, or horizons, that appear different from one another (Fig. 2.3). The horizons within a soil profile vary in thickness depending on the intensity of the soil-forming factors, although their boundaries are not always easy to distinguish. Uppermost layers of mineral soils are the most altered, whereas the deeper layers are most similar to the original parent material. Alterations of the solum, the parent material most affected during soil formation, involve (1) decay of organic matter from plant residues and roots and accumulating as dark-colored humus (the organic matter-enriched horizons nearest the soil surface are called A horizons); (2) eluviation by water of soluble and colloidal inorganic and organic constituents from surface soils to varying depths in the profile; and (3) accumulation of inorganic and organic precipitates in subsurface layers. These underlying, enriched layers are referred to as B horizons. The C horizons are the least weathered of the mineral soil profile. Organic soils are commonly saturated with water and mainly consist of mosses, sedges, or other hydrophytic vegetation, with the upper material being referred to as the O layer. In upland forested areas where drainage is better, folic-derived organic materials accumulate as an L-F-H layer.

The vadose zone is the underlying, unsaturated, parent material extending downward from the soil surface to where it reaches the water table and the soil becomes saturated. Below the solum, this zone contains relatively unweathered parent material, low in organic matter and nutrients, and intermittently deficient in O<sub>2</sub>. The thickness of the vadose zone can fluctuate considerably during the season, depending on soil texture, soil water content, and height of the soil water table. When the water table is near the surface, for example as in wetlands, it may be narrow or nonexistent. But in arid or semiarid areas, where soils are well drained, the vadose zone can extend for several meters.

## **III PHYSICAL ASPECTS OF SOIL**

Dimensions of the features, commonly encountered when considering the soil habitat, range from a few meters (the pedon, the fundamental unit of soil within the landscape), down through a few millimeters (soil aggregates and the fine earth fraction), to a few micrometers (living microorganisms and clay minerals) and nanometers (fragments of microbial cell walls).



**FIG. 2.3** Profiles of common mineral and organic soils: Mollisol (top left), Spodosol (top right), Oxisol (bottom left), and Histisol (bottom right).

## A Soil Texture

The larger mineral particles include stones, gravels, sands, and coarse silts that are generally derived from ground-up rock and mineral fragments. Although particles  $> 2$  mm in diameter may affect the physical attributes of a soil, it is

the fine earth fraction, those individual mineral particles  $\leq 2$  mm in diameter, that describe soil texture. The fine earth fraction of soil particles ranges in size over four orders of magnitude: from 2.0 mm to smaller than 0.002 mm in diameter. Sand-sized particles are individually large enough (2.0 to 0.05 mm) to be seen by the naked eye and feel gritty when rubbed between the fingers in a moist state. Somewhat smaller, silt-sized particles (0.05 to 0.002 mm) are microscopic and feel smooth and slippery even when wet. Clay-sized particles are the smallest of the mineral particles ( $<0.002$  mm), seen only with the aid of a microscope, and when wet, they form a sticky mass. The proportions of sand, silt, and clay are referred to as soil texture, and terms such as *sandy loam*, *silty clay*, and *clay loam* are textural classes used to identify the soil's texture. When investigating a field site, considerable insight to the behavior and properties of the soil can be inferred from its texture (e.g., soil water characteristics, nutrient retention, susceptibility to compaction); thus, it is often one of the first properties to be measured (How to determine soil texture by feel; <http://www.youtube.com/watch?v¼GWZwbVJCNe>; see online supplemental material at: <http://booksite.elsevier.com/9780124159556>).

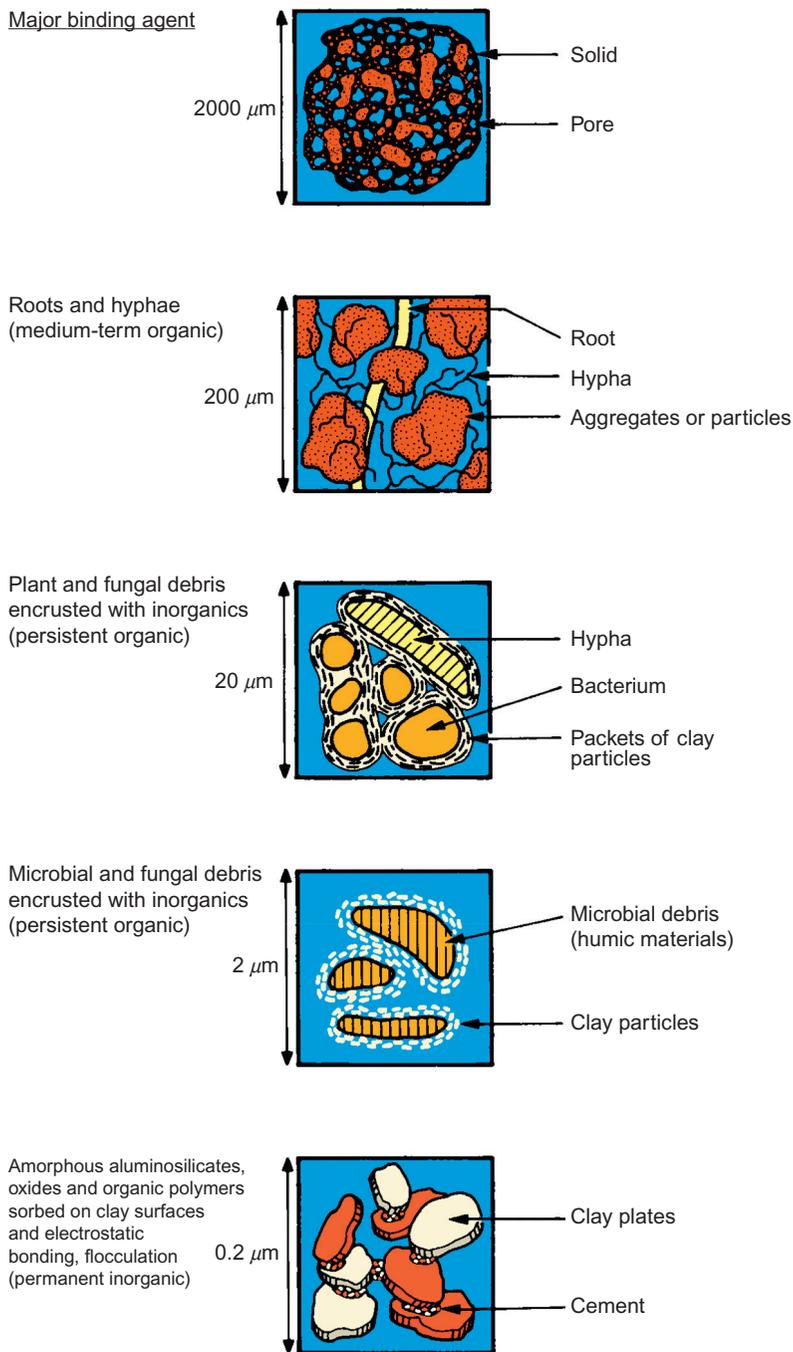
The surface of mineral soils contains an accumulation of living biomass and dead and decomposing organic material. The SOM typically accounts for 1-10% of the total soil mass, and most is intimately associated with the mineral fraction, especially the surfaces of clays, making it difficult to isolate from the soil for study. The larger, recognizable remains of plant, animal, and soil organisms that can be separated from soils by hand picking, sieving techniques, and floatation are referred to as particulate organic matter or light fraction organic matter. These tissues undergo continuous microbial decay and turnover and, over periods of years to decades, accumulate in soils as brown to black-colored, chemically complex colloids.

## **B Aggregation of Soil Mineral Particles**

Typically the individual mineral particles in surface soils are coated and become glued together with organic matter and by inorganic cements, forming spatial clusters within the soil structural matrix known as aggregates or peds. The arrangement of aggregates is referred to as soil structure and, together with soil texture, is of particular importance in regulating biotic activity because of its influence on water content and aeration.

SOM and clay minerals play particularly important roles in aggregate formation due to their large surface area and negative electrostatic charge (i.e., are colloidal in nature). Loamy or clayey soils are usually strongly aggregated, whereas sandy and silty soils are weakly aggregated.

Tisdall and Oades (1982) presented a conceptual model of the hierarchical nature of soil aggregation, which described the linkages between the architecture of the soil habitat and the role of microbial activity in its genesis (Fig. 2.4). Three assemblages of aggregates are generally recognized with



**FIG. 2.4** Model of the aggregated, hierarchical nature of the soil system and major binding agents (from Tisdall and Oades, 1982).

diameter classes of: 0.002-0.020 mm, 0.020-0.250 mm, and > 0.250 mm and are referred to as microaggregates, mesoaggregates, and macroaggregates, respectively.

Microaggregates are formed by flocculation of fine silt and clay particles, amorphous minerals (composed of oxides and hydroxides of aluminum, silicon, iron, and manganese, and silicates of aluminum and iron), and nonhumic and humic substances, which are largely dominated by electrostatic and van der Waals forces. Polyvalent cations, such as  $\text{Al}^{3+}$ ,  $\text{Fe}^{3+}$ ,  $\text{Ca}^{2+}$ , and  $\text{Mg}^{2+}$  adsorbing onto their surfaces and reacting with exposed functional groups, promote these flocculation reactions. Sticky polysaccharides and proteins, derived from plant and animal tissues, microbial cells and exudates from roots, hyphae, and bacteria further enhance these stabilization reactions. In particular, wherever intense biological activity associated with organic matter decomposition occurs, the extensive exopolysaccharides microorganisms produce glue minerals and organic particles that form microaggregates. The core of mesoaggregates is usually the residual debris left from the decay of plant and microbial tissues. Bits of this decaying particulate organic matter, and their colonizing microbial biofilms are encrusted with fine mineral particles, which act as nuclei for the formation of larger aggregates. In the rhizosphere, hyphae of arbuscular mycorrhizal fungi (AMF) contribute to aggregation.

Macroaggregates are formed where a network of living plant roots and root hairs, fungal hyphae, and fibrous organic matter physically enmesh clusters of micro- and mesoaggregates for a period sufficient for them to be chemically linked. Casts deposited by earthworms and fecal pellets left by microarthropods, mites, and colembola also contribute to aggregate formation. This hierarchical organization of aggregate formation (i.e., large aggregates being composed of smaller aggregates), which in turn are composed of even smaller aggregates, is characteristic of most undisturbed surface soils.

Micro- and mesoaggregates tend to be especially resistant to mechanical breakdown, for example, from the impact of rainfall, or from slaking-rapid rewetting of dry soil, or from freezing and thawing. The restricted size of the pores contained within these aggregates can limit accessibility of the associated colloidal humus to microbial decay and restrict interactions of soil organisms, thereby protecting microorganisms from predation by fauna. Macroaggregates usually remain intact as long as the soil is not disturbed, for example, by earthworm and other faunal activity, by the impact of intense rainfall, or by tillage machinery. The pore spaces contained within macroaggregates, referred to as intraaggregate pore space, are important for providing soil aeration and water retention for plant growth. The pore space surrounding macroaggregates, collectively referred to as the interaggregate pore space, is where plant roots and larger fragments of plant residues are found. Macroaggregation is important for controlling biotic activity and SOM turnover in surface soils because the interaggregate pore space allows for the exchange of oxygen and other gases with the atmosphere (Ball, 2013). It may also regulate the accessibility of particulate organic matter to decay

by soil organisms. By determining the nature and pore-size distribution within soil, macroaggregation can give fine-textured, clayey- and loamy-textured soils the beneficial pore space characteristics for aeration, water infiltration, and drainage of sandy soils. (Method for evaluation of aggregation in the field shown in Fig. 2.7: <http://www.sac.ac.uk/vess>; see online supplemental material at: <http://booksite.elsevier.com/9780124159556>); Guimarães et al., 2011.

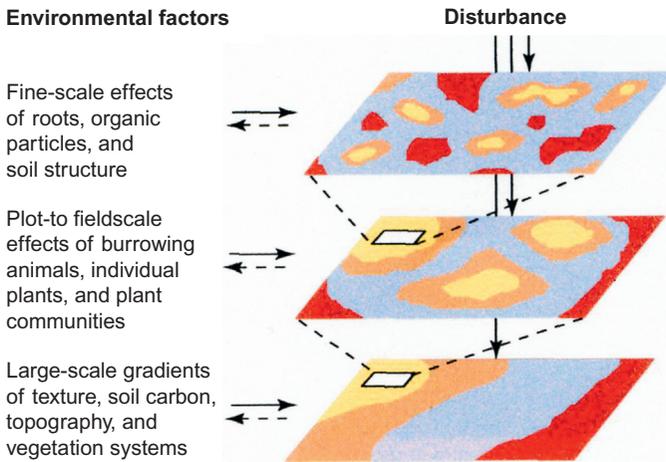
## IV SOIL HABITAT AND SCALE OF OBSERVATION

### A Scale of Soil Habitat

The habitat provided by the soil is characterized by heterogeneity, measured across scales from nanometers to kilometers (Chapter 1), differing in chemical, physical, and biological characteristics in both space and time. At various levels within this continuum of scales, different soil properties used to characterize the habitat can assume greater or lesser importance, depending on the function or attribute that is under consideration. For higher organisms, such as animals that range over wide territories, the habitat may be on the scale of a landscape and beyond. For studies of climate change effects on soil respiration, the distribution of hydrologic features within a watershed affecting soil warming may be appropriate. At the other extreme, evaluations of more specific processes that impact individual microbial species' functioning (e.g., denitrifier activity, and oxygen and substrate availability) may be possible only at the microhabitat scale. The habitat of a particular soil organism includes the physical and chemical attributes of its location and also the biological component of the habitat that influences the growth, activities, interactions, and survival of other organisms associated with this space (i.e., other microorganisms, fauna, plants, and animals). The spatial attributes across all scales must be considered when describing soil organism activity. Soil habitat spatial heterogeneity is an important contributor to the coexistence of species in soil microbial communities, thereby enhancing overall soil biodiversity by promoting the persistence of individual populations. It is central to the explanation of the high species-richness of the soil population.

Studies have confirmed that soil organisms are usually not randomly distributed, but exhibit predictable spatial patterns over wide spatial scales (Fig. 2.5). Spatial patterns of soil biota also affect the spatial patterns of activity and the processes they carry out. As an example, inorganic N production from ammonification of SOM can accumulate if the microbial processes of mineralization and immobilization are physically separated in soil space. This also occurs where plant residues are left on the soil surface compared to when they are incorporated into the soil.

Although the main factors influencing the gross behavior of soil organisms are known, the role of spatial distribution has not been studied in detail. There are few methods currently available that enable the study of microbial activity



**FIG. 2.5** Determinants of spatial heterogeneity of soil organisms. Spatial heterogeneity in soil organism distributions occurs on nested scales and is shaped by a spatial hierarchy of environmental factors, intrinsic population processes, and disturbance. Disturbance operates at all spatial scales and can be a key driver of spatial heterogeneity, for example, through biomass reduction of dominant organisms or alteration of the physical structure of the soil substrate. Feedback between spatial patterns of soil biotic activity and heterogeneity of environmental factors adds further complexity (dotted arrows) (from *Ettema and Wardle, 2002*).

*in situ* at the level of the soil microhabitat. It is common practice for soil scientists, after collecting samples in the field from a soil profile, to pass the soil through a 2-mm sieve to remove the plant debris and macrofauna and then homogenize the samples before analysis.

## B Pore Space

On a volume basis, mineral soils are about 35-55% pore space, whereas organic soils are 80-90% pore space. Total soil pore space can vary widely for a variety of reasons, including soil mineralogy, bulk density, organic matter content, and disturbance. Pore space can range from as low as 25% for compacted subsoils in the lower vadose zone to more than 60% in well-aggregated clay-textured surface soils. Even though sand-textured soils have a higher mean pore size, they tend to have less total pore space than do clay soils.

Soil pore space is defined as the percentage of the total soil volume occupied by soil pores:

$$\% \text{pore space} = [\text{pore volume} / \text{soil volume}] \times 100 \quad (2.2)$$

Direct measurements of soil pore volume are difficult to perform, but estimations can be obtained from data on soil bulk density and soil particle density, using the following formulas:

$$\text{Soil bulk density : } D_b (\text{Mg m}^{-3}) = \text{soil mass (Mg)} / \text{soil bulk volume (m}^3) \quad (2.3)$$

$$\text{Soil particle density : } D_p (\text{Mg m}^{-3}) = \text{soil mass (Mg)} / \text{soil particle volume (m}^3) \quad (2.4)$$

(assumed to be  $2.65 \text{ Mg m}^{-3}$  for silicate minerals, but can be as high as  $3.25 \text{ Mg m}^{-3}$  for iron-rich tropical soils and as low as  $1.3 \text{ Mg m}^{-3}$  for volcanic soils and organic soils)

$$\% \text{pore space} = 100 - [(D_b/D_p) \times 100] \quad (2.5)$$

where:

$$D_b = \text{soil bulk density, Mg m}^{-3} \quad (2.6)$$

$$D_p = \text{soil particle density, Mg m}^{-3} \quad (2.7)$$

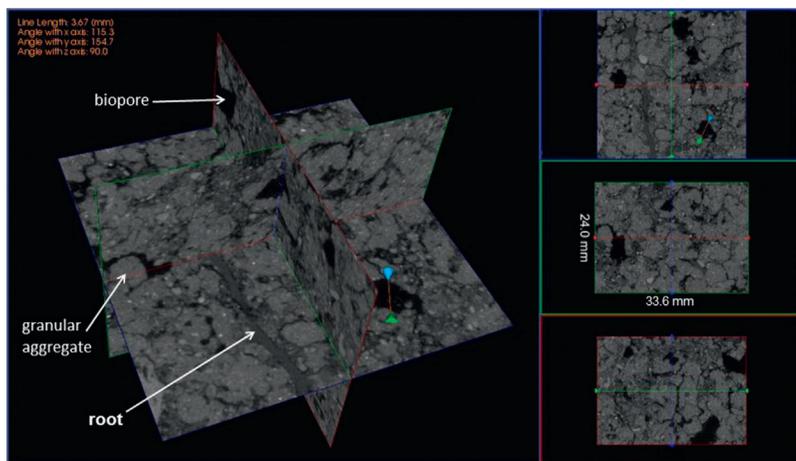
Although total pore space is important, the size and interconnection of the pores are the key in determining the habitability of the soil. Total pore space is usually divided into two size classes, macropores and micropores, largely based on their ability to retain water left after drainage under the influence of gravity (see soil water content). Macropores are  $\geq 10 \mu\text{m}$  in diameter and allow rapid diffusion of air and rapid water infiltration and drainage. They can occur as the spaces between individual sand and coarse silt grains in coarse-textured soils and in the interaggregate pore space of well-aggregated loam- and clay-textured soils. Macropores can be created by roots, earthworms, and other soil organisms forming a special type of pore termed biopore. Biopores are typically lined with organic matter and clay and are ideal habitats for soil biota. They provide continuous channels often extending from the soil surface throughout the soil for lengths of one or more meters.

Soil pores  $< 10 \mu\text{m}$  in diameter are referred to as micropores and are important for water retention for plants and for providing an aqueous habitat for microorganisms. Water flow and gaseous diffusion in micropores are slow, but help to provide a stable, local environment for soil biota. Although the larger micropores, together with the smaller macropores, can accommodate plant root hairs and microorganisms, pores  $\leq 5 \mu\text{m}$  in diameter are not habitable by most microorganisms. They may even restrict diffusion of exoenzymes and nutrients, thereby inhibiting the uptake of otherwise potential substrates. Although surface soils are typically about 50% pore space on a volume basis, only a quarter to a half of this pore space may be habitable by soil microorganisms due to restricted pore size (Table 2.1).

**TABLE 2.1** Relationship of soil water potential and equivalent pore diameter to soil pore space (% of total soil porosity) across a textural gradient

General pore categories	Soil water potential (kPa)	Equivalent pore diameter ( $\mu\text{m}$ )	Sandy loam	Loam	Clay loam
Macropores (aeration)	0	All pores filled (saturated)	100	100	100
	-1	$\leq 300$	5	3	2
	-10	$\leq 30$	24	14	12
Micropores (capillary)	-33	$\leq 10$	8	5	3
	-100	$\leq 3$	3	14	7
	-1000	$\leq 0.3$	2	6	17
	-10,000	$\leq 0.03$	2	7	12
		Total pore space (%)	44	49	53

The work of Kubiena in the 1930s contributed significantly to our understanding of the nature of soil solid and pore space at the microscopic scale. Much of this early research was based on the examination of thin sections (25  $\mu\text{m}$  thick) of intact blocks of soil. Adaptation of advancements in data acquisition and computer-assisted analysis of digital imagery during the past quarter century have led to the quantitative spatial analysis of soil components. Recent developments in microcomputerized X-ray tomography (CT scanning) allow for the study of the properties of the soil's intact three-dimensional structure. These systems have high resolution capabilities (10-30  $\mu\text{m}$ ), which allow differentiation of solids for quantifying the distribution of organic and mineral materials. The technology is also able to readily distinguish air-filled and water-filled pore space. However, it is still not possible to distinguish microorganisms from soil particles with this technology. A CT image of a soil core in three orthogonal planes is shown in Fig. 2.6. Highly attenuating features, such as iron oxide nodules, appear bright in the imagery, whereas features with low attenuation capability, such as pore space, appear dark. Though distinguishing microorganisms from soil particles with this technology is still limited, newer technologies, such as secondary ion mass spectrometry (SIMS), have the potential to do so ([www.sciencemag.org/cgi/content/full/304/5677/1634/DC1](http://www.sciencemag.org/cgi/content/full/304/5677/1634/DC1); Movies S1 and S2; see online supplemental material at <http://booksite.elsevier.com/9780124159556>).



**FIG. 2.6** X-ray CT image of forest mull Ah horizon from Nepean, Ontario, Canada. The three images on the right correspond to the orthogonal planes in the main image. Voxel size of imagery is 40  $\mu\text{m}$ ; dimensions of the full image are 33.6 mm (width)  $\times$  33.6 mm (length)  $\times$  24.0 mm (height). By convention, air-filled pore space is dark, and solid materials are lighter in tone. Diameter of the measured biopore, an earthworm channel, is  $\sim 3.7$  mm.

## V SOIL SOLUTION CHEMISTRY

An understanding of the chemistry of the soil solution, providing an environment for soil organisms, needs to take into account the nature and quantity of its major components: water, dissolved organic matter and inorganic constituents, and  $\text{O}_2$  and  $\text{CO}_2$ . The biogeochemistry of the soil solution is mainly determined by acid-base and redox reactions (Chapter 9). The thermodynamic activities of protons and electrons in soil solution define the chemical environment that controls biotic activity. Conceptually, both can be considered as flowing from regions of high concentration to regions of low concentration, with soil microbial activity having a profound effect on regulating this flow.

### A Soil pH

Protons supplied to the soil from atmospheric and organic sources react with bases contained in aluminosilicates, carbonates, and other mineralogical and humic constituents. In a humid climate with excess precipitation, and given sufficient time, basic cations ( $\text{Na}^+$ ,  $\text{K}^+$ ,  $\text{Ca}^{2+}$ ,  $\text{Mg}^{2+}$ ) will be exchanged from mineral and organic constituents by  $\text{H}^+$  and be leached from the surface soil. The presence of calcite and clay minerals, such as smectites, which are saturated with basic cations, retards the rate of acidification. Continued hydrolysis results in the formation of the secondary minerals (kaolinite, gibbsite, and goethite) and a soil solution buffered between pH 3.5 and 5. Semi-arid and arid conditions lead to an opposite trend: a soil solution buffered at an alkaline pH. Soil pH influences

a number of factors affecting microbial activity, such as solubility and ionization of inorganic and organic soil solution constituents, which in turn affect soil enzyme activity. There are large numbers of both organic and inorganic acids found in soils, although the majority of these acids are relatively weak.

Measurements of soil solution pH provide important data for predicting potential microbial reactions and enzyme activity in soil. Though easily measured in a soil paste with a pH electrode, interpretation of its effects on microbial processes can be difficult. This is largely because concentrations of cations sorbed to the surfaces of negatively charged soil colloids are 10-100 times higher than those of the soil solution. Thus, if enzymes are sorbed to colloid surfaces, their apparent pH optimum would be 1-2 pH units higher than if they were not sorbed. An example of this is soil urease activity, which has an apparent pH optimum of 8.5-9.0 in soil that is about 2 pH units higher than optimal urease activity measured in solution.

## **B Soil Redox**

The most reduced material in the biosphere is the organic matter contained in living biomass. Organic matter in soils ranges from total dominance, as in peatlands, to the minor amounts found in young soils or at depth in the vadose zone. The metabolic activity of soil organisms produces electrons during the oxidation of organic matter, which then must be transferred to an electron acceptor, the largest of which is O<sub>2</sub> contained in freely drained, aerobic soils. The O<sub>2</sub> contained in soil air or present in soil solution can be consumed within hours depending on the activity of soil organisms and is replenished by O<sub>2</sub> diffusion. If O<sub>2</sub> consumption rates by soil organisms are high, due to an abundant supply of readily decomposable organic C, or if O<sub>2</sub> diffusion into the soil is impeded because of waterlogging or restricted pore sizes, due to clay texture or soil compaction, soil solution O<sub>2</sub> concentrations continue to decrease. When all available dissolved O<sub>2</sub> has been used, the solution changes from aerobic (oxic) to anaerobic (anoxic). Microbial activity will then be controlled by the movement of electrons to alternative electron acceptors.

Development of anaerobic conditions results in a shift in the activity of the soil microbial populations. The activity of aerobic and facultative organisms, which dominate well-drained oxic soils, decreases while the activity of obligate anaerobic organisms and fermentative organisms increases. This switch in electron acceptors promotes the reduction of several important elements in soil, including N, Mn, Fe, and S in a process known as anaerobic respiration and of CO<sub>2</sub> by methanogenesis.

Redox potential ( $E_H$ ) measurements provide an indication of the soil aeration status. They are a measure of electron availability occurring as a result of electron transfer between oxidized (chemical species that have lost electrons) and reduced (chemical species that have gained electrons) chemical species. The measurements are often used to predict the most probable products of

biological reactions. For example,  $N_2O$  can be produced from nitrification under aerobic conditions and from denitrification under moderately reducing conditions where the reduction intensity is not strong enough to completely reduce nitrate to  $N_2$  gas.

The magnitude of  $E_H$  depends on  $E^\circ$ , and also on the relative activities of the oxidant and the reductant. These various quantities are related by the Nernst equation,

$$E_H(V) = E^\circ - (0.0591/n) \log(\text{reduct})/(\text{oxid}) + (0.0591 m/n)\text{pH} \quad (2.8)$$

The  $E_H$  is the electrode potential of the standard hydrogen electrode,  $E^\circ$  is the standard half-cell potential,  $n$  is the number of electrons transferred,  $m$  is the number of protons exchanged, *reduct* is the activity of the reduced species, and *oxid* is the activity of the oxidized species.

The major redox reactions occurring in soils and the electrode potentials for these transformations are shown in Table 2.2. Typically, dissolved  $O_2$  and  $NO_3^-$  serve as electron acceptors at  $E_H \sim 350$  to  $400$  mV and above. When their concentrations in soil solution become low, Mn, Fe, and  $SO_4^{2-}$  serve as alternate electron acceptors. Their reduction processes occur over a wider range compared with  $O_2$  and  $NO_3^-$  reduction and methane production. Mn and Fe serve as electron acceptors at  $\sim 350$  mV for Mn and at  $250$  mV to  $\sim 100$  mV for Fe. When  $Fe^{3+}$  in Fe oxide is reduced, the oxide dissolves and  $Fe^{2+}$  goes into solution. Sulfate reduction occurs at an  $E_H$  as high as  $350$  mV and to as low as  $\sim 100$  mV. Methane production begins when  $E_H$  is close to  $\sim 100$  mV.

Although the activity of electrons can be described by pE,  $E_H$  has the advantage of being a standard measurement for investigations of soil redox potential both in the laboratory and the field. Soil  $E_H$  can be obtained relatively easily from measurements of the pore water using a platinum (Pt)

**TABLE 2.2** Important Redox Pairs and the Approximate  $E_H$  Values at the Occurrence of Transitions at the Reference Soil pH of 7.0

	Oxidized Form	Reduced Form	Approximate $E_H$ at Transformations (mV)
Oxygen	$O_2$	$H_2O$	+600 to +400
Nitrogen	$NO_3^-$	$N_2O, N_2, NH_4^+$	250
Manganese	$Mn^{4+}$	$Mn^{2+}$	225
Iron	$Fe^{3+}$	$Fe^{2+}$	+100 to -100
Sulfur	$SO_4^{2-}$	$S^{2-}$	-100 to -200
Carbon	$CO_2$	$CH_4$	Less than -200

electrode. However, the values for soil  $E_H$  can be difficult to interpret as the Pt electrode measurement does not reflect changes in all the chemical species involved in redox reactions and also responds to changes in pH. Often two or more redox reactions occur simultaneously, and thus measured  $E_H$  usually reflects a mixed potential.

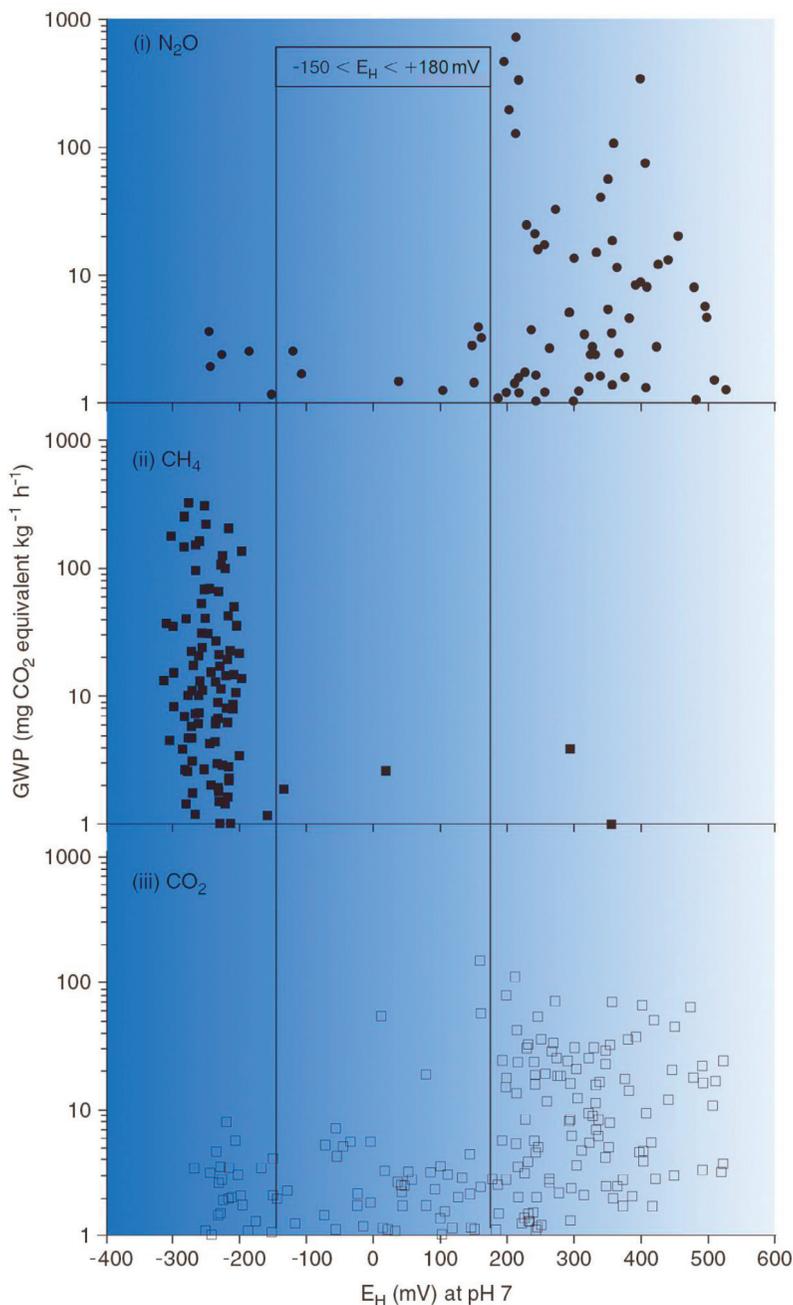
Platinum-electrode  $E_H$  measurements are still useful and can be interpreted as a semiquantitative assessment of a soil's redox status. In studies of paddy soils, for rice production,  $E_H$  measurements can be used to monitor progressive development of reducing conditions and can distinguish oxic and anoxic conditions. These fields provide a unique environment for studying the relationship between soil  $E_H$  and greenhouse gas emissions because of controlled irrigation and drainage practices (Fig. 2.7). During the flood season, the paddy soils are a major source of  $\text{CH}_4$  and an important source of  $\text{N}_2\text{O}$  when they are drained. Strategies designed to mitigate  $\text{CH}_4$  emissions from submerged rice fields can adversely affect greenhouse warming potential by stimulating higher  $\text{N}_2\text{O}$  emissions. The different  $E_H$  conditions required for  $\text{N}_2\text{O}$  and  $\text{CH}_4$  formations and the trade-off pattern of their emissions as found in rice fields makes it a challenge to abate the production of one gas without enhancing the production of the other. An  $E_H$  greater than  $-150$  mV, but less than  $+180$  mV, offers the minimum global warming potential contribution from these rice soils.

## C Soil Aeration

Molecular diffusion dominates the transport of gases in the soil. Diffusion through the air-filled pores maintains the gaseous exchange between the atmosphere and the soil, and diffusion through water films of varying thickness maintains the exchange of gases with soil organisms. Diffusivity through both pathways can be described by Fick's law:

$$J = D \, dc/dx, \quad (2.9)$$

where  $J$  is the rate of gas diffusion ( $\text{g cm}^{-2} \text{sec}^{-1}$ ),  $D$  is the diffusion coefficient for soil air and for water ( $\text{cm}^2 \text{sec}^{-1}$ ),  $c$  is the gas concentration ( $\text{g cm}^{-3}$ ),  $x$  is the distance (cm), and  $dc/dx$  is the concentration gradient. The gaseous diffusion coefficient in soil air is much smaller than that in the atmosphere. The limited fraction of total pore volume is occupied by continuous air-filled pores, pore tortuosity, and soil particles. Water reduces the cross-sectional area and increases the mean path length available for diffusion. For soil air, this is referred to as the effective diffusion coefficient,  $D_e$ , and is a function of the air-filled porosity. Likewise, tortuosity due to particulate material present in soil solution reduces rates of gaseous diffusion. As shown in Table 2.3, diffusion of gases in water is  $\sim 1/10,000$  of that in air. Gaseous diffusion through a  $10\text{-}\mu\text{m}$  water film would take the same time as diffusion through a  $10\text{-cm}$  air-filled pore.



**FIG. 2.7** Global warming potential (GWP) contribution of  $\text{N}_2\text{O}$ ,  $\text{CH}_4$ , and  $\text{CO}_2$  as a function of soil  $E_H$ . All eight soils showed the same pattern of (i)  $\text{N}_2\text{O}$ , (ii)  $\text{CH}_4$ , and (iii)  $\text{CO}_2$  dynamics with soil  $E_H$  change from high to low. The figure is plotted on a logarithmic scale to cover a wide range of values. Global warming potential contributions below  $1 \text{ mg CO}_2$  equivalent  $\text{kg}^{-1} \text{h}^{-1}$  were considered insignificant and not illustrated in the figure for clarity. (from *Kewei and Patrick, 2004*).

**TABLE 2.3** Temperature Effects on Gaseous Diffusion in (A) Air, in (B) Water, and (C) Gas Solubility in Water

Temperature (°C)	N <sub>2</sub>	O <sub>2</sub>	CO <sub>2</sub>	N <sub>2</sub> O
<b>A. Gaseous Diffusion Coefficients in Air (cm<sup>2</sup>/sec)</b>				
0	0.148	0.178	0.139	0.179
10	0.157	0.189	0.150	0.190
20	0.170	0.205	0.161	0.206
30	0.180	0.217	0.172	0.218
<b>B. Gaseous diffusion Coefficients in Water (×10<sup>-4</sup> cm<sup>2</sup>/sec)</b>				
0	0.091	0.110	0.088	0.111
10	0.130	0.157	0.125	0.158
20	0.175	0.210	0.167	0.211
30	0.228	0.275	0.219	0.276
<b>C. Solubility Coefficients (volume of dissolved gas relative to volume of water, cm<sup>3</sup>/cm<sup>3</sup>)</b>				
0	0.0235	0.0489	1.713	1.30
10	0.0186	0.0380	1.194	1.01
20	0.0154	0.0310	0.878	0.71
30	0.0134	0.0261	0.665	0.42

## VI ENVIRONMENTAL FACTORS, TEMPERATURE, AND MOISTURE INTERACTIONS

### A Soil Temperature

The physical, chemical, and biological processes that occur in soil are influenced by temperature in vastly different ways and are the most important environmental factors that must be considered. Whereas rates of molecular diffusion always increase with increasing temperature, solubility of gases in soil solution does not and can even decrease.

The relation between a chemical reaction rate and temperature was first proposed by Arrhenius:

$$k = A e^{-E_a/RT} \quad (2.10)$$

The constant  $A$  is called the frequency factor and is related to the frequency of molecular collisions;  $E_a$  is the activation energy required to initiate the reaction;  $R$  is the universal gas constant and has a value of  $8.314 \text{ J mol}^{-1} \text{ T}^{-1}$ ;  $e$  is the

base of the natural logarithm;  $T$  is the absolute temperature (in  $^{\circ}\text{K}$ ); and  $k$  is the specific reaction rate constant ( $\text{time}^{-1}$ ).

Conversion of Eq. (2.10) to natural logarithmic form gives

$$\ln k = (-E_a/RT) + \ln A \quad (2.11)$$

By determining values for  $k$  over a moderate range of soil temperatures, a plot of  $\ln k$  versus  $1/T$  results in a linear relationship, providing the activation energy is constant over the temperature interval;  $E_a$  is obtained from the slope of the line and  $A$  from the  $Y$  intercept.

A similar equation can be used to describe temperature effects on enzyme activity:

$$k_{(\text{cat})} = \kappa(k_B T/h)e^{-\Delta G^\ddagger/RT} \quad (2.12)$$

where  $k_{(\text{cat})}$  is the reaction rate ( $\text{time}^{-1}$ ),  $\kappa$  is the transmission coefficient,  $k_B$  is the Boltzmann constant,  $h$  is the Planck constant,  $\Delta G^\ddagger$  is the activation energy,  $R$  is the universal gas constant, and  $T$  is the absolute temperature (in  $^{\circ}\text{K}$ ). The transmission coefficient varies significantly with the viscosity of the soil solution, which increases by a factor of almost 2 over a temperature drop from  $20^{\circ}\text{C}$  to near  $0^{\circ}\text{C}$ . In both instances, a change in temperature results in an exponential change in the reaction rate, the magnitude of which is a function of the activation energy. Note also that there is an inverse relationship between reaction rate and activation energy.

The relationship between temperature and biologically mediated processes is complicated as individual species differ in their optimal temperature response, different microbial communities are active as temperatures change, and microorganisms are able to adapt by altering their physiology and cellular mechanisms, membrane fluidity and permeability, and structural flexibility of enzymes and proteins.

The relative sensitivity of soil microbial activity to temperature can be expressed as a  $Q_{10}$  function, which is the proportional change in activity associated with a  $10^{\circ}\text{C}$  temperature change

$$Q_{10} = (k_2/k_1)^{[10/(T_2-T_1)]} \quad (2.13)$$

where  $k_2$  and  $k_1$  are the rate constants for a microbial process under study at temperatures differing by  $10^{\circ}\text{C}$ . It is generally accepted that a  $Q_{10}$  of  $\sim 2$  can be used to describe the temperature sensitivity of soil biochemical processes, such as soil respiration, over the mesophilic temperature range ( $20$ – $45^{\circ}\text{C}$ ); that is, microbial activity at  $30^{\circ}\text{C}$  is twofold higher than it is at  $20^{\circ}\text{C}$ . At temperatures beyond  $45^{\circ}\text{C}$ , the microbial community composition shifts from mesophilic to thermophilic, and microorganisms adapt by increasing concentrations of saturated fatty acids in their cytoplasmic membrane and by production of heat-stable proteins.

Microorganisms that have an upper growth temperature limit of  $\leq 20^{\circ}\text{C}$ , commonly referred to as psychrophiles, are capable of growth at low temperatures by

adjusting upward both the osmotic concentration of their cytoplasmic constituents to permit cell interiors to remain unfrozen and the proportion of unsaturated fatty acids in their cytoplasmic membrane. A common adaptive feature of psychrophiles to low temperatures is that their enzymes have much lower activation energies and much higher (up to 10-fold) specific activities than do those of mesophiles, resulting in a reaction rate,  $k_{(cat)}$ , that is largely independent of temperature. Although microbial activity slows at lower temperatures, rates are significantly higher and more sensitive to temperature changes than those predicted from studies over the mesophilic temperature range. As an example, researchers studying decomposition of SOM, soil respiration, and N mineralization have reported values for  $Q_{10}$  increasing to near 8-10 with soil temperatures approaching 0°C (Kirschbaum, 2013). Díaz-Raviña et al. (1994) reported  $Q_{10}$  values between 3.7 and 6.7 for the 0-10°C interval for thymidine incorporation and between 5.0 and 13.9 for acetate incorporation for a soil bacterial community.

Low temperatures are common over vast areas of the earth and include soils in temperate, polar, and alpine regions where mean annual temperatures are < 5°C. Soils in these environments contain a great diversity of cold-adapted microorganisms, able to thrive even at subzero temperatures and to survive repeated freeze/thaw events. A bacterium isolated from a permafrost soil in northern Canada can grow at -15°C and remain metabolically active at temperatures down to -25°C. Microbial activity during cold periods when plants are dormant or soils are barren can play a significant role in overwinter losses of soil nutrients, in particular for N with overwinter leaching and by denitrification during freeze/thaw.

Very few soils maintain a uniform temperature in their upper layers. Variations may be either seasonal or diurnal. Because of the high specific heat of water, wet soils are less subject to large diurnal temperature fluctuations than are dry soils. Among factors affecting the rate of soil warming, the intensity and reflectance of solar irradiation are critical. The soil's aspects (south- versus north-facing slopes), steepness of slope, degree of shading, and surface cover (vegetation, litter, mulches) determine effective solar irradiation. Given the importance of soil temperature in controlling soil processes, models of energy movement into the surface soil profile have been developed. They are based on physical laws of soil heat transport and thermal diffusivity and include empirical parameters related to the temporal (seasonal) and sinusoidal variations in the diurnal pattern of near-surface air temperatures. The amplitude of the diurnal soil temperature variation is greatly dampened with profile depth.

## **B Soil Water**

Soil water affects the moisture available to organisms as well as soil aeration status, the nature and amount of soluble materials, the osmotic pressure, and the pH of the soil solution. Water acts physically as an agent of transport by mass flow and as a medium through which reactants diffuse to and from sites of reaction. Chemically, water acts as a solvent and as a reactant in important

chemical and biological reactions. Of special significance in the soil system, and to microbial cells in particular, is the fact that water adsorbs strongly to itself and to surfaces of soil particles by hydrogen bonding and dipole interactions. Soil water content can be measured on a mass or volume basis. Gravimetric soil water content is the mass of water in the soil, measured as the mass loss in a soil dried at 105°C (oven-dry weight) and is expressed per unit mass of oven-dried soil. Volumetric soil water content is the volume of water per unit volume of soil. Soil water is also described in terms of its potential free energy, based on the concept of matric, osmotic, and gravitational forces affecting water potential. Soil water potential is expressed in units of pascals (Pa), or more commonly, kilopascals (kPa), with pure water having a potential of 0. Matric forces are attributed to the adhesive or adsorption forces of water attraction to surfaces of mineral and organic particles and to cohesive forces or attraction to itself. These forces reduce the free energy status of the water. Solutes dissolved in soil solution also contribute to a reduction in the free energy of water and give rise to an osmotic potential that also is negative. Combined, the matric and osmotic forces are responsible for the retention of water in soils. They act against gravitational forces that tend to draw water downward and out of the soil.

When the gravitational forces that drain water downward are exactly counterbalanced by the matric and osmotic forces that hold onto the water, the soil is said to be at field capacity, or at its water-holding capacity. This will occur after irrigation, after a heavy rainfall, or after spring thaw, which leave the soil saturated with a soil water potential = 0 kPa. Gravitational forces begin to immediately drain away water in excess of what can be retained by matric + osmotic forces, leaving the soil after one to two days at field capacity. By definition, the field capacity for loam and clay loam soils is a soil water potential of -33 kPa, and for sandy soils it is -10 kPa. (How soil properties affect soil water storage and movement: <http://www.youtube.com/watch?v=jWwtDKT6NAw>; see online supplemental material at: <http://booksite.elsevier.com/9780124159556>).

Water retention, or soil water content at a given soil water potential, is a function of the size of pores present in the soil, or pore size distribution (Table 2.1). Soils of different texture have very different water contents, even though they have the same water potential. Surface tension is an important property of water influencing its behavior in soil pores. In addition, due to water's strong cohesive forces, it has a high surface tension. Based on matric forces and properties of surface tension, the maximum diameter of pores filled with water at a given soil water potential can be estimated using the Young-Laplace equation:

$$\frac{\text{Maximum pore diameter } (\mu\text{m})}{\text{Retaining water}} = \frac{300}{\text{Soil water potential (kPa)}} \quad (2.14)$$

Those soil pores  $\geq 10 \mu\text{m}$  in diameter drain under the influence of gravitation forces, given that the soil water potential at field capacity is  $-33 \text{ kPa}$ .

Soil water characteristics are difficult to assess in laboratory and field studies. A program developed by Solomon and Rauls can be used to estimate soil water characteristics based on soil texture and organic matter content, which are commonly measured physical and chemical soil properties (<http://hydrolab.arsusda.gov/soilwater/Index.htm>; see online supplemental material at: <http://booksite.elsevier.com/9780124159556>). Soil water potential determines the energy that an organism must expend to obtain water from the soil solution. Generally, aerobic microbial activity in soil is considered optimal over soil water potentials ranging from about  $-50$  to  $-150 \text{ kPa}$ , which is 30-50% of the soil's total pore space, depending on its texture and bulk density (Table 2.1). Aerobic activity decreases as soil becomes wetter and eventually saturated, due to restricted  $\text{O}_2$  diffusion. When greater than 60% of the pore space is water-filled, activity of microorganisms able to use alternative electron acceptors increases (e.g., that of anaerobic denitrifiers).

As the soil dries and water potential decreases, water films on soil particles become thinner and more disconnected, restricting substrate and nutrient diffusion, and increasing the concentration of salts in the soil solution. Although many plants grown for agricultural purposes wilt permanently when the soil water potential reaches  $-1500 \text{ kPa}$ , rates of soil microbial activity are less affected as the relative humidity within the soil remains high. Respiration rates, for example, can still be  $\sim 90\%$  of maximum at this low soil water potential. Studies both in laboratory incubations and in the field have reported that the logarithm of soil matric potential is a good predictor of the effect of water content on soil respiration.

Although some microorganisms are able to adapt to low soil water potentials by accumulating osmolytes (amino acids and polyols) or changing the properties of their outer membrane, rapid changes in soil water potential associated with drying/rewetting cause microbes to undergo osmotic shock and induce cell lysis. A flush of activity by the remaining microbes, known as the Birch effect, results from mineralization of the labile cell constituents released.

Different communities of organisms are active over the range of water potentials commonly found in soils. Protozoa are active at water potentials near field capacity in water films  $\geq 5 \mu\text{m}$  thick, whereas microorganisms can be active at lower water potentials, due to their size and association with the surfaces of soil particles. Even though fungi are generally considered to be more tolerant of lower soil water potentials than are bacteria, presumably because soil bacteria are relatively immobile and rely more on diffusion processes for nutrition, the opposite has also been shown to be true due to differences in community structure. Table 2.4 shows differences in the ability of various soil organisms to tolerate water stress. Sulfur and ammonium oxidizers, typified by *Nitrosomonas* and *Thiobacillus* species, respectively, are less tolerant of water stress than are the ammonifiers and typified by

**TABLE 2.4** Ability of Different Organisms to Tolerate Water Stress

Soil Water Potential (kPa)	Water Activity ( $A_w$ )	Organism
-0.5	> 0.99	Protozoa
-1,500	0.99	<i>Rhizobium</i> , <i>Nitrosomonas</i>
-4,000	0.97	<i>Bacillus</i>
-10,000	0.93	<i>Clostridium</i> , <i>Fusarium</i> , <i>Eurotium</i>
-25,000	0.83	<i>Micrococcus</i> , <i>Penicillium</i> , <i>Aspergillus</i>
-65,000	0.62	<i>Xeromyces</i> , <i>Chrysosporium</i> , <i>Monascus</i>

*Clostridium* and *Penicillium*. Ammonium may accumulate in droughty soils at the water potentials where ammonifiers are still active but restrict nitrification.

Soil moisture and temperature are the critical factors affected by climate regulating soil biological activity. This control is affected by changes in the underlying rates of enzyme-catalyzed reactions and sizes of the substrate organic and inorganic pools. Where water is nonlimiting, biological activity may depend primarily on temperature. Standard Arrhenius theory can be used to predict these temperature effects. However, as soils dry, moisture is a greater determining factor of biological processes than is temperature. Likely these two environmental influences do not affect microbial activity in linear fashion, but display complex, nonlinear interactions that reflect the individual responses of the various microorganisms and their associated enzyme systems.

The interactions of temperature, moisture, and organisms are exemplified by the current concerns about climate change on soil biology. A hundred years ago, Swedish scientist Svante Arrhenius asked the important question, "Is the mean temperature of the ground in any way influenced by the presence of the heat-absorbing gases in the atmosphere?" He went on to become the first person to investigate the effect that doubling atmospheric  $\text{CO}_2$  would have on global climate. The question was debated throughout the early part of the 20th century and is a main concern of Earth systems scientists today. The earth's surface is warming for several reasons, one of which is increased emissions of greenhouse gases from soils due to past and current agricultural management practices, deforestation, combustion of fossil fuels, and industrial pollution. Global temperatures have increased by  $\sim 0.5^\circ\text{C}$  over the past 100 years and are expected to increase by  $1^\circ\text{C}$  to  $6^\circ\text{C}$  by 2100. Although this represents only a few degrees temperature change, global warming will dramatically increase microbial decay rates of the organic matter stored in the boreal forests and tundra regions, estimated to contain  $\sim 30\%$  of the global soil C (Kirschbaum, 1995). Global

warming is already melting glaciers and ice sheets at accelerated rates, and the permafrost thaw rate has more than tripled over the past half century.

The critical concern is that SOM decomposition is stimulated to a greater extent than is net plant productivity, the C input to SOM. Theory also suggests that the more resistant organic matter compounds, with high activation energies, would become more decomposable at higher temperatures (Davidson and Janssens, 2006). Degradation of cellulose, hemicellulose, and other components of SOM by extracellular enzymes is the rate-limiting step in CO<sub>2</sub> emissions. However, feedback mechanisms characteristic of all biogeochemical cycles may dampen effects of temperature changes. Soils are complex, with interactions such as change in rainfall patterns that affect plant productivity, soil water storage, and nutrient cycling. Many of the environmental constraints affect decomposition reactions by altering organic matter (substrate) concentrations at the site at which all decomposition occurs, that of the enzyme reaction site. We must also consider decomposition rates at the enzyme affinity level; Michaelis-Menten models of enzyme kinetics are covered in Chapter 16 and energy yields in Chapter 9. The kinetic and thermodynamic properties of extracellular enzymes and their responses to environmental factors are now being considered in models of the effects of global warming on carbon cycling. Changes in microbial community structure (Chapter 8) will also have profound influences. The goal of this chapter is to provide an environmental boundary of the soil habitat and a description of its fundamental physical and chemical properties. With this as a foundation, later chapters in this volume explore in detail information about organisms, their biochemistry, and interactions.

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