similar to sea breezes; air flows toward and up the mountain slope during the day and downward at night. Drainage winds occur from the gravitation of cold air off high ground in intermountain areas. This air seeps down the slopes and gathers in valleys to produce a gentle or moderate cool or cold breeze. Where mountains are located close to a coast, drainage winds can be severe. An example is the Santa Ana wind, which flows down the Santa Ana Canyon of southern California and spreads over the lowlands toward the coast. Chinook winds are strong, dry, warm winds that develop along the lee slopes of mountain ranges. In the United States, lowlands just east of the Rocky Mountains are subjected to chinook winds from westerly airflow across the mountains.

Wind Effects on Turf

Disruption by winds of the boundary layer surrounding turfgrass leaves and wind-induced soil deposition have already been discussed. Wind can be very beneficial to turfgrasses or highly detrimental, depending on its intensity. Air flowing at a few miles per hour can accelerate heat transfer to substantially cool a turf during hot weather. The importance of wind in drying turfgrass foliage and, thus, reducing disease incidence is most evident on sites where dense trees and shrubs obstruct air movement. The incidence of brown patch, *Pythium,* and *Helminthosporium* diseases is usually much higher on these sites. Exposed sites at relatively high elevations, however, may be subjected to severe winds that promote rapid drying of the turf. Such sites may require more irrigation during the summer. In winter, severe winds across turfgrass sites without snow cover can cause substantial desiccation and loss of turf. Protective windbreaks, including trees, mounds, and other landscape features, can effectively reduce the potential for desiccation injury to turf.

Wind is also important in disseminating weed seeds and vegetative propagules, fungal spores, salt sprays, and nearly all other foreign substances and organisms that can adversely affect turf. Atmospheric pollutants from industrial and other sources are also carried by wind. Sulfur dioxide, fluorides, ozone, and some nitrogen-containing gases can be transported in sufficient concentrations to directly injure turfgrasses or to weaken them so that they are less resistant to other environmental stresses.

EDAPHIC ENVIRONMENT

The science that deals with the influence of soil and other media on the growth of plants is called *edaphology.* The edaphic environment of a turfgrass community may be composed of synthetic materials, native soil, organic residues, or any combination thereof. Where combined, these media may form a uniform mixture or occur as distinct layers within the turf soil profile. Naturally occurring soil profiles have more or less distinct layers, or horizons, resulting from various activities termed soil-forming processes (Figure 4.15). The surface soil (A horizon) is a zone that has received leachates from organic materials situated at its surface and from which various substances have been leached by percolating water. It is typically higher in organic matter and more favorably structured than the lower soil horizons. Below is the subsoil, or B horizon, where accumulation of

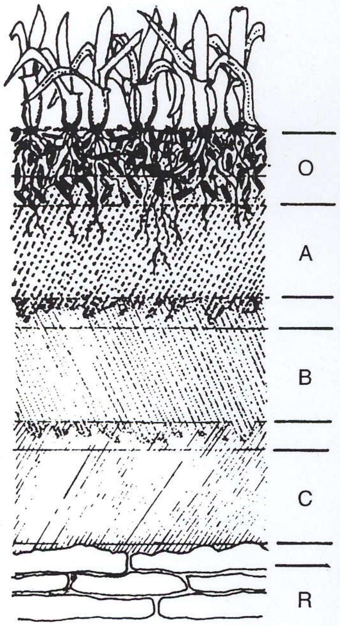


Figure 4.15. Undisturbed soil profile under turf showing surface organic residue (O), topsoil (A), subsoil (B), substratum (C), and bedrock, if present (R).

leachates has occurred. Subsoil is a far less suitable medium for supporting plant growth than is soil from the A horizon. The C horizon is a zone of partially weathered parent material resulting from the deposition of transported materials or from decomposition of underlying bedrock. Thousands of years of weathering, leaching, and organic matter accumulation are necessary to transform parent materials into soil.

On turfgrass sites, the zonation of the soil profile may be obscured by mixing or transfer of materials from different horizons, especially where deep excavations of soil have been made during construction operations. Also, the planned incorporation of various amendments may have substantially altered the characteristics of a naturally formed soil. Because soil formation requires many thousands of years, productive soils must be regarded as limited environmental resources requiring careful and intelligent management.

Soil is important as a plant growth medium in terms of its fertility, water relations, gas exchange, and physical support of plant roots. For turfgrass plantings, the soil and plant community should also provide a firm but resilient surface that resists compaction from traffic and use.

If a block of turf including soil several inches deep were examined, three distinct phases of physical composition would be evident: solid, liquid, and air. Soil, organic matter in different stages of decomposition, and roots and lateral shoots form a solid phase. Organic matter may occur as indistinguishable additions to the soil mineral matter or as

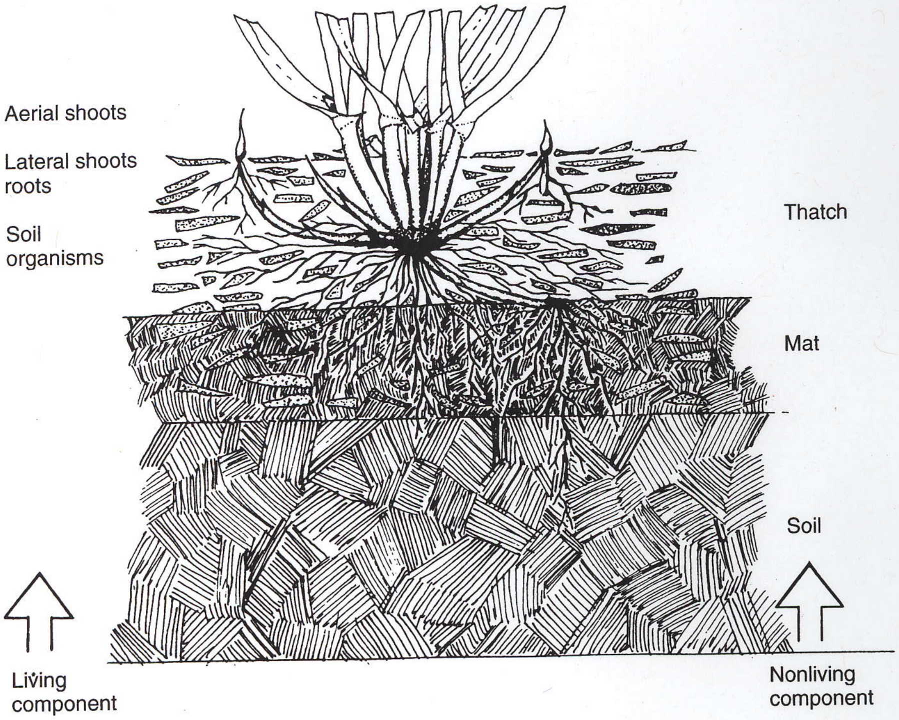


Figure 4.16. A block of turf with thatch, mat, and soil layers in the profile.

organic strands positioned within or above the surface soil. The undecomposed or partially decomposed layer of organic matter situated above the soil surface is called *thatch.* When integrated with the surface soil, it forms a thatchlike derivative called *mat.* A specific block of turf may contain one or more of these edaphic features (thatch, mat, and soil), depending on naturally occurring phenomena, cultural practices, and turfgrass genotype (Figure 4.16). Thatch and mat formation will be discussed further in a later section of this chapter.

Pore spaces of varying sizes and shapes permeate the soil and organic materials. If the turf block were completely saturated, all pores would be filled with water and dissolved substances (liquid phase). If it were allowed to dry, all but the smallest pores would be filled with air (air phase).

The relative proportions of the three compositional phases within the total volume of the block largely determine the suitability of the soil as a growth medium for turfgrasses. Generally, a composition of one-half solids, one-fourth liquid, and one-fourth air is considered an optimum proportion for a plant growth medium. The capacity of a growth medium to sustain this proportion is a function of numerous physical, chemical, and biological factors within the edaphic environment. Therefore, turfgrass edaphology involves the physics, chemistry, and biology of soil media.

Table 4.1. Textural Classification of Soil Particles (USDA)

Separate	Diameter size (mm)
Very coarse sand	2.00–1.00
Coarse sand	1.00–0.50
Medium sand	0.50–0.25
Fine sand	0.25–0.10
Very fine sand	0.10–0.05
Silt	0.05–0.002
Clay	< 0.002

Soil Physics

The physical properties of a soil, including texture and structure, directly affect its aeration, moisture, and temperature, and indirectly affect fertility and the activity of soil organisms.

Texture

Soil texture is determined by the size of soil particles and their relative proportion. Soil particles range in size from less than 1 micrometer (μm, micron) to 2 millimeters (2 mm or 2000 υm) in diameter. Larger particles may occur in soil, but these are considered separately as gravel or rocks. Depending on their size, soil particles are grouped into various soil separates, including clay, silt, and sand. Further divisions are recognized to separate sand into five textural groups. Therefore, there are a total of seven soil separates according to the USDA system of classification (Table 4.1). The proportion of sand, silt, and clay determines the textural class. There are twelve textural classes, as illustrated in the textural triangle (Figure 4.17).

Associated with texture is the amount of surface area that exists within a given weight of soil. One gram of very coarse sand has approximately 11 cm^2 of surface, while the same weight of clay may have a total surface area of 8 million cm^2. Clay is important in chemical reactions involving the adsorption and exchange of plant nutrients. The pore sizes within a predominantly clay soil are so small, however, that much of the water contained in them is generally unavailable to plants. Silt pores are larger and retain higher amounts of plant-available moisture, while sand pores are so large that they contribute little to water retention. Sand is important, however, in promoting soil aeration and drainage.

Structure

Soil structure refers to the arrangement of soil particles. Clay forms aggregates in which individual particles are held together in various configurations. An individual aggregate

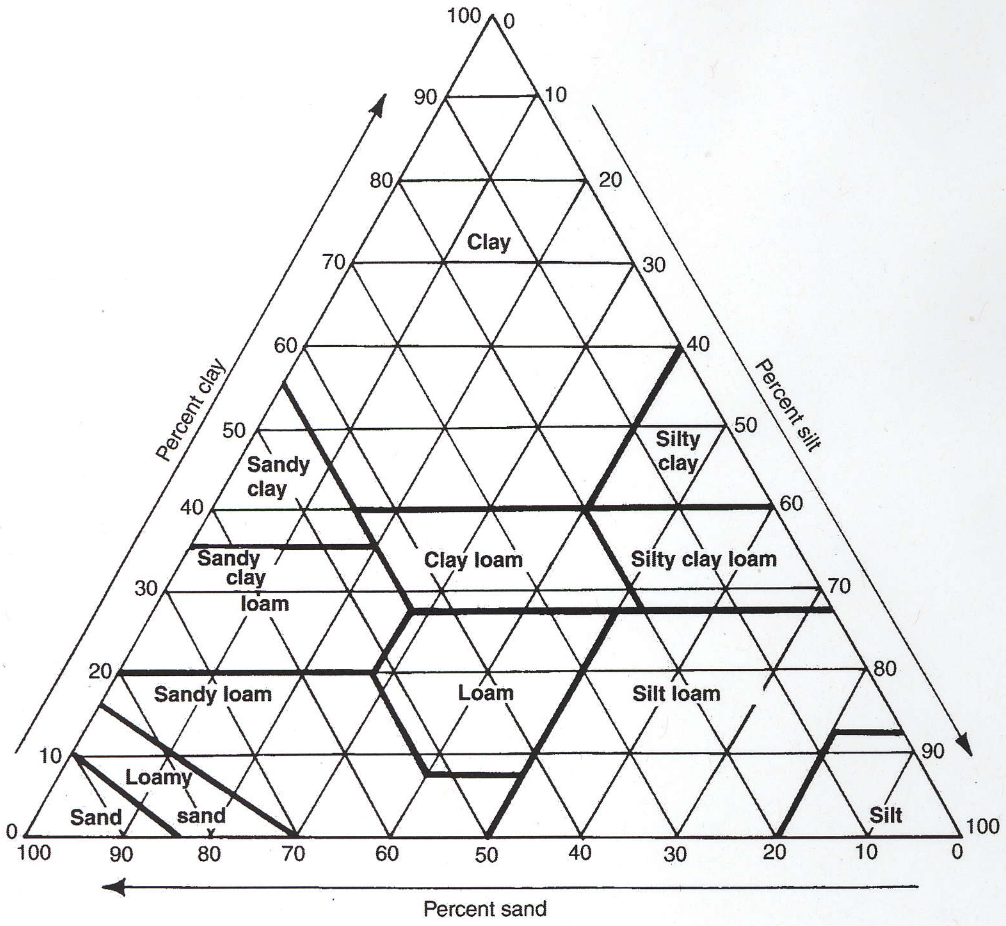


Figure 4.17. Textural triangle with soil textural classes reflecting relative percentages of sand, silt, and clay.

may be as large as or larger than a sand particle. Thus, aggregation provides a medium in which a broad range of pore sizes exists. If soil aggregates are sufficiently stable, a well-aggregated clay soil can serve as an excellent medium for plant growth. However, aggregates differ in their structural stability depending on the specific clay minerals present and the strength of the binding agents holding particles together. Traffic, splashing rain or irrigation, and cultivation also tend to destroy soil structure. Usually, the best soils for plant growth are those that contain a reasonable proportion of various soil separates and possess a favorable soil structure that has been promoted and sustained through proper cultural practices.

The first step in aggregate formation is the bringing together of soil particles by various forces, including freezing and thawing, wetting and drying, root growth, and the activities of soil organisms. Once joined, polyvalent cations (Ca^{2+}, Mg^{2+}, Al^{3+}, and others) can bridge adjacent soil particles by forming electrostatic bonds. Various cementing agents resulting from decomposing organic matter or hydroxylation of cations can also bind soil particles together. Without structure, a fine-textured soil would be a dense plastic mass when wet, and bricklike when dry. If crushed, a dry structureless clay will be flourlike in appearance.

Density

Density is the mass of substance per unit volume, usually expressed as grams per cubic centimeter (g/cc). Water has a density of 1 g/cc (at 4°C and 760 mm Hg). Two density measurements, particle density and bulk density, are commonly used for soils. Particle density (PD) is the density of dry, solid soil particles. It averages 2.65 g/cc. Bulk density (BD) is the dry weight of an undisturbed volume of soil. A well-structured, fine- or medium-textured soil will have an abundance of large and small pores, and its bulk density will be low compared to the same soil in a compacted state. If a soil could be so compacted that virtually all pore spaces were removed, bulk density would equal particle density. Comparisons of the bulk densities of different soils may not provide reliable indices for determining their suitability for growing plants. Sands have predominantly large pores, yet their bulk densities are high because of the relatively large mass of solids making up these media. For example, a bulk density of 1.5 g/cc may indicate a rather compacted loam soil with insufficient large pores for drainage. However, in a coarse sand of the same bulk density, aeration would not be a limiting factor in plant growth, but water retention might be. Within soil type, bulk density values are valuable for assessing the physical condition of a soil. Soil porosity can also be determined from bulk density:

$$\%\ \text{porosity} = 100 - (\text{BD}/\text{PD} \times 100)$$

A loam soil with a bulk density of 1.3 would, therefore, have 49% pore spaces:

$$100 - (1.30/2.65 \times 100) = 49\%$$

The percent pore space is only a measure of total porosity and does not directly indicate the distribution of different-sized pores. Within a particular soil type, however, the percent porosity can provide a reasonable assessment of pore size distribution if a sufficient bank of information exists to compare plant growth response to different soil porosities.

Moisture

The importance of soil structure and density lies in their influence on the number and size of pores and the consequent movement of water and air within the soil. Water drains rapidly from large (aeration-type) *macropores* due to gravitational force. This is called

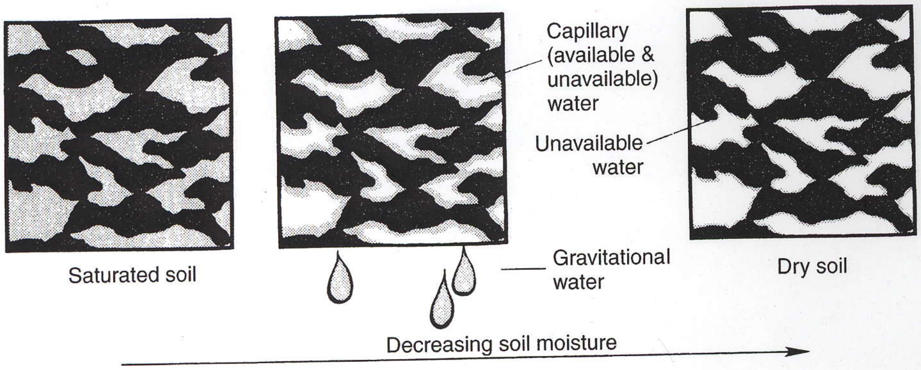


Figure 4.18. Illustration of different types of soil water. Gravitational water drains from the large (aeration) pores, while capillary water is held as films on soil surfaces.

gravitational water (Figure 4.18). The remaining water is retained as thin films on the surfaces of soil particles and as wedges where two or more particles come together. To absorb this water, plants must overcome forces of adhesion and cohesion that hold water in the soil. Adhesion is the attraction between soil surfaces and water; cohesion is the attraction between water molecules. The portion of retained water that plant roots can absorb is called *available water,* and is contained within the *mesopores.* When water films are reduced in thickness, the attraction of the soil for water becomes greater, and eventually plants can no longer secure enough water to satisfy their needs. The tightly held water that is essentially unavailable to plants is simply called *unavailable water;* it is contained within the *micropores.*

A well-structured soil will release enough water in response to gravity so that aeration porosity is adequate to sustain healthy plants. The water content of the soil following drainage of the aeration pores is called *field capacity,* and the air content is called *aeration capacity.* These are not precise values, but are useful as indications of soil structure for a given soil. In a poorly structured soil, removed water may not be replaced by air because of soil shrinkage. Such soils form massive clods with large cracks instead of friable granules.

The thickness of a water film at which plants can no longer absorb sufficient water to sustain growth is about the same for all soil types. However, the water content, measured in grams of water per gram of dry soil, at which this occurs varies among soils because of large differences in total surface area. When plants growing in a particular soil wilt irreversibly, the soil is said to be at its *permanent wilting point,* and the amount of water remaining in the soil is called the *permanent wilting percentage.* This amount ranges from 1% to 2% for sandy soils to 25% to 30% for fine-textured (clayey) soils. Therefore, a measure of soil water content cannot serve as a guide to a plant's soil-water requirement, but a good relationship exists between a plant's water needs and the work required to remove a unit of water from the soil. An expression of the energy status of water is called the *water potential* and is symbolized by Ψ_W, the Greek letter psi. Water potential is usually measured in units of pressure (e.g., 1 atmosphere = 14.7 pounds per square inch [psi] = 760 mm Hg = 1.013 bar or 101.13 centibars [cb] = 101.13 kilopascals [kPa] or 0.1013 megapascals [MPa]). Relative to pure free water, which has a water potential of 0, the water potential of the soil solution (Ψ_{SW}) is almost always negative. Soil water potential can be dissected into individual components, usually written as the formula

$$\Psi_{SW} = \Psi_M + \Psi_O + \Psi_P$$

in which soil water potential equals the sum of the matric potential (Ψ_M), osmotic potential (Ψ_O), and pressure potential (Ψ_P). The matric potential (Ψ_M) accounts for the reduction in free energy of water when it exists as films or layers adsorbed onto soil surfaces due to the attraction between the surface and the water. As adsorbed water is under tension (i.e., negative pressure), its energy is less than that of pure free water; thus, its matric potential is always negative. As a soil dries, the films of moisture adhering to particle surfaces become thinner, and the matric potential, as well as the soil water potential, are correspondingly reduced. At field capacity the matric potential is typically between -10 and -33 kPa; at the permanent wilting point, it declines to -1500 kPa. The osmotic potential (Ψ_O) represents the effects of dissolved solutes on soil water potential. Solutes such as nitrate ions reduce the free energy of water due to the attraction between the solute and the water; thus, as with matric potential, the osmotic potential of soil water is always negative. As the concentration of dissolved salts and other materials in the soil solution increases, the osmotic potential as well as the soil water potential are correspondingly reduced. Finally, the pressure potential (Ψ_P) accounts for the effect of water under pressure. Where a water table exists, the water at the bottom of the table may be under positive pressure; however, in unsaturated soils there is no liquid pressure, $\Psi_P = 0$ and the soil water potential is simply the sum of the matric and osmotic potentials ($\Psi_{SW} = \Psi_M + \Psi_O$). Because the salt concentration of the soil solution is inversely proportional to the amount of water present, as a soil containing an abundance of soluble salts dries, the soil water potential can decline precipitously. For example, where $\Psi_O = -216$ kP and $\Psi_M = -200$ kP, $\Psi_{SW} = -416$ kP, which indicates a major reduction in soil water availability.

Differences in soil water potential between two locations in a soil, called the *water potential gradient,* provide a force that causes water to flow from locations of higher to lower water potential. The greater the water potential gradient the stronger the force moving water through the soil. But soil water movement is influenced by more than just the soil water potential gradient. *Soil water conductivity* is a measure of the ease with which the soil conducts water. In compacted soils, for example, conductivity is low because of high resistance to flow. Conversely, well-structured soils conduct water more rapidly. Where a plant is pulling water from the soil immediately surrounding its roots, replenishment of this root-zone soil moisture depends on (1) the water potential gradient between the root-zone soil and a location in the soil where water is available and (2) the conductivity of the soil.

In a saturated soil, water potential near the soil surface approaches zero. As the soil drains, water potential becomes progressively lower (more negative). Water movement in saturated and unsaturated soils should be considered as two distinct processes. Saturated flow occurs when all or most of the pores are filled with water. It takes place through large pores and, therefore, is most rapid in coarse-textured soils. The principal force acting on the water is gravity, and the direction of flow is primarily downward. If the number of large pores decreases suddenly at a given soil depth, downward movement is restricted and water may accumulate above the interface where the two media meet, resulting in a temporary water table. This occurs in sand or thatch overlying loam soil or with a compacted subsoil.

Unsaturated water flow occurs in soils in which the large pores are not filled with water. The rate of unsaturated flow depends on the thickness of water films surrounding soil particles; thicker water films allow faster flow rates than thinner films due to the differences in water potential. Thus, water moves faster in moist soils than in dry soils. Unsaturated flow proceeds in any direction, irrespective of gravitational force. The "wick action" or capillary flow of water from lower to upper soil locations is actually unsaturated flow.

Where the continuity of water films is disrupted, as at the interface between a fine-textured soil and an underlying coarse-textured soil, unsaturated flow is slowed or may stop altogether (Figure 4.19). This can result in an accumulation of water, called a *perched water table,* above the interface. Water will not move across the interface until the water potential in the above soil builds to a level sufficient to overcome the attraction between the water and the fine-textured soil. When sufficient water potential has built up from continued downward flow toward the interface, water will enter the coarse-textured soil and be conducted away. Where a coarse-textured soil is underlain by a fine-textured soil, downward flow through the soil profile will slow to a rate determined by the hydraulic conductivity of the fine-textured soil. If that rate is slower than the rate at which water is entering the coarse-textured soil, a *temporary water table* may form as long as water continues to enter the soil at the same rate and persist until the fine-textured soil absorbs all of the free water situated above it.

This principle can be applied to turf soils in which layers exist either by design or by error. Consider a fine-textured soil that has been modified by incorporation of sand or other coarse amendments. If the incorporation is not uniform but results in subsurface layers of coarse material, water flow within the profile can be disrupted. On the other hand, where a layer of coarse sand has been intentionally positioned beneath finer-textured sand with some loam soil and organic amendments, a perched water table will develop following irrigation or rainfall. In this instance, the perched water table is desired to compensate for the low water retention of the fine sand. A USGA green is constructed in this fashion to combine the advantages of compaction resistance and moisture retention in the root zone. Results are only satisfactory, however, where the design includes a critical depth of the surface medium; a too-shallow surface layer will not drain properly, while a layer that is too deep will be too dry at the surface. This can be illustrated using a rectangular household sponge measuring 5 by 3 by 1 inch (Figure 4.20). When the sponge is saturated and positioned with its 5-inch and 3-inch sides in the horizontal plane, very little water drains out. Turning the sponge 90° to position the 3-inch side vertically results in more drainage. Rotating the sponge so that its 5-inch side is vertical results in still further drainage. This demonstration shows the relationship between height of the

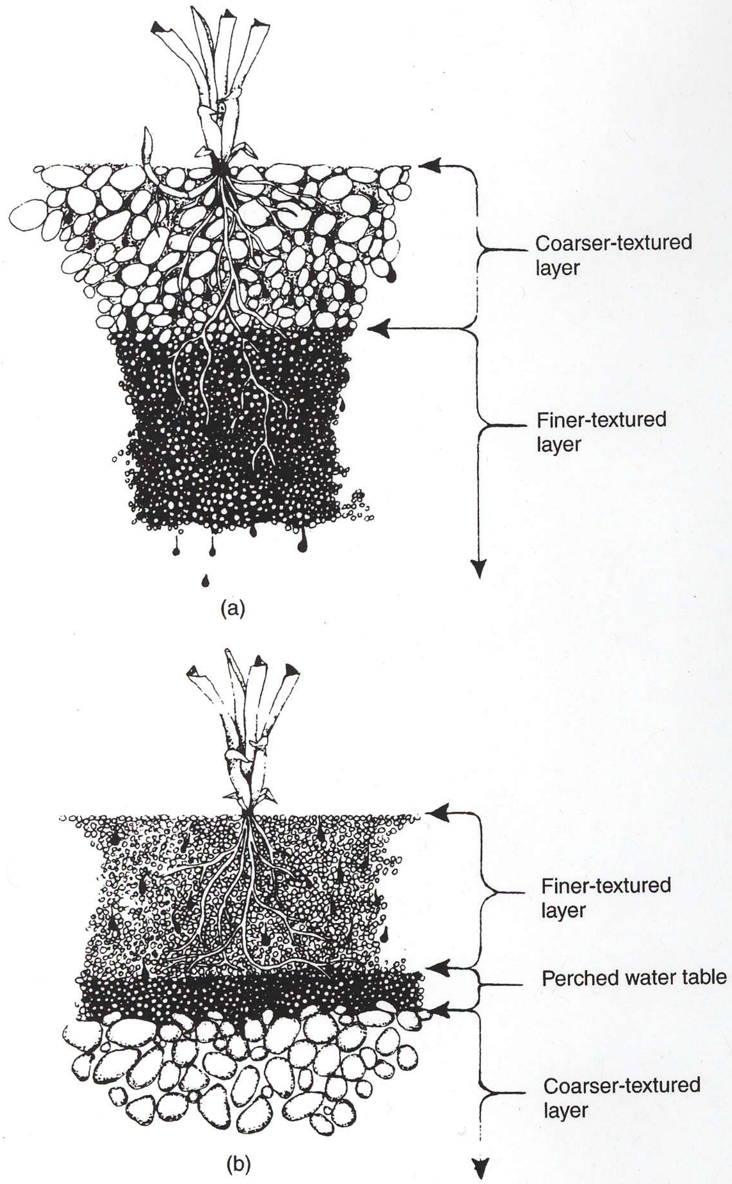


Figure 4.19. Layered turf-soil profiles in which a coarser-textured layer (thatch, sand) overlies a finer-textured soil (a), and in which a finer-textured soil overlies a layer of coarser texture (b), resulting in a perched water table.

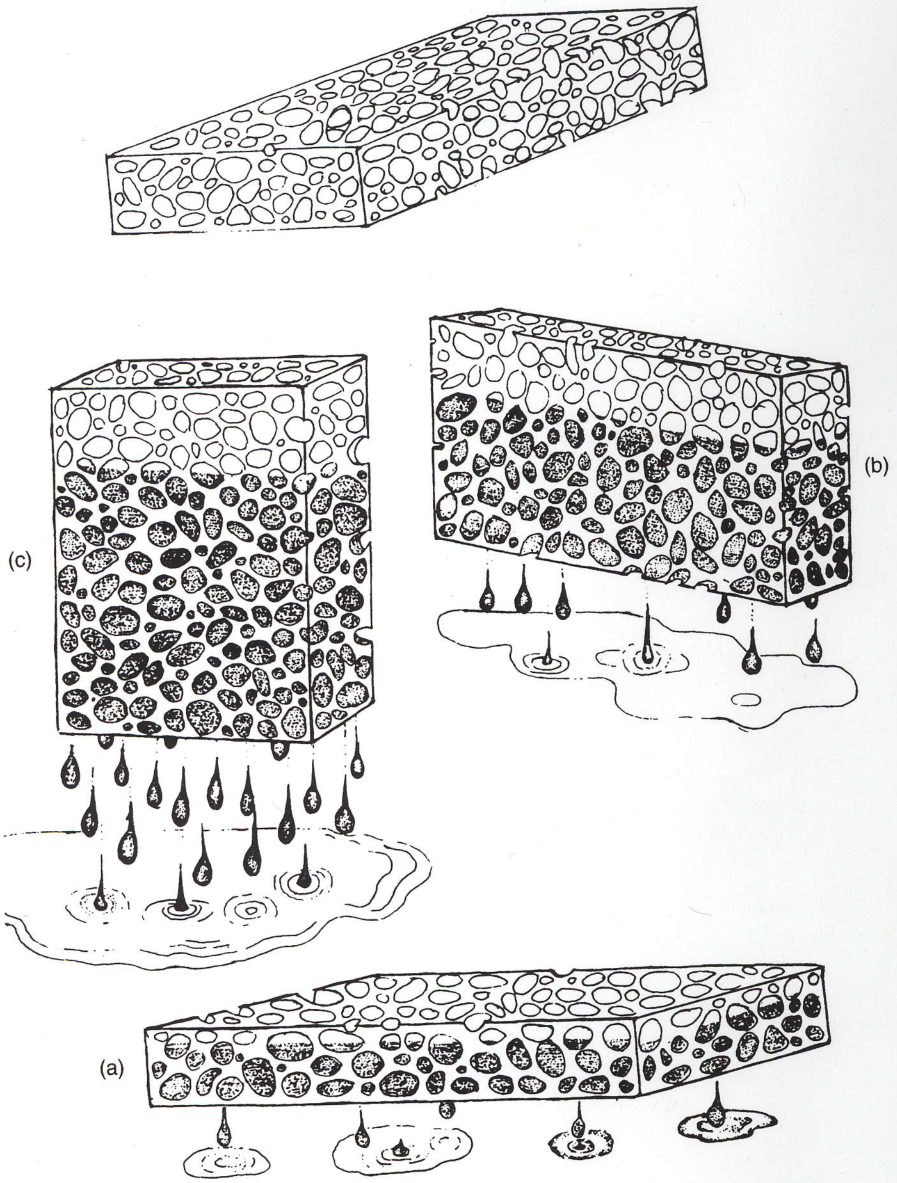


Figure 4.20. Water drainage from a saturated sponge (a) with the thin side oriented vertically, (b) after being rotated 90°, and (c) after its longest side is oriented vertically.

water column and water retention within the pore volume of the sponge. Cutting the sponge horizontally into three equal sections while it is still in the 5-inch vertical orientation and squeezing each section to remove internal moisture would reveal that the uppermost section was driest and the lowermost section was wettest.

The sponge analogy facilitates understanding of some soil water phenomena. In a USGA green, the sponge represents the surface soil medium, and the air or "free space" beneath the sponge represents the coarse sand or gravel layer underlying the surface medium. A workable greens design depends on a rather precise application of principles of soil physics. The settled depth (depth after soil settling) of the surface medium must be correct, and the physical composition of the medium must be such that air and moisture within the turfgrass root zone are sufficient to sustain growth. The USGA green and other designs are discussed further in Chapter 9.

Aeration

The process by which soil air is replaced by atmospheric air is called *soil aeration.* Soil air has higher concentrations of carbon dioxide and water vapor but less oxygen than atmospheric air due to the consumption of oxygen and production of carbon dioxide by soil organisms. The magnitude of difference depends on the rate of gaseous exchange between the atmosphere and the soil. Aeration is brought about by processes of diffusion and mass flow.

Diffusion is the movement of gases through air-filled pores from regions of higher to lower gas concentration and is proportional to air-filled porosity. Diffusion is low in compacted soils because of reduced pore size and number and the discontinuity of soil pores. Similarly, diffusion is low in wet soils because of the absence or reduction of air-filled pores and the extremely low diffusion rate of air in water. Mass flow occurs as a result of the following:

1. Expansion and contraction of soil gases due to temperature and barometric pressure changes.
2. Soil air removal through precipitation and irrigation, and replacement as water is removed by drainage, plant use, and evaporation.
3. Wind action causing air to be forced into the soil at some locations and pulled out at others. Relative to diffusion, mass flow is considered to have a minor influence on soil aeration.

Poorly aerated soils are often deficient in oxygen. Oxygen is used by plant roots and soil organisms for respiration, which produces carbon dioxide. Without an adequate exchange of gases between the atmosphere and the soil, oxygen levels decline and carbon dioxide levels increase in soil air. This can result in reduced absorption of nutrients and water by plant roots, since they must have sufficient oxygen for respiration to generate the energy necessary for this process. Microbial decomposition of organic matter is also inhibited in oxygen-deficient soils, as is the bacterial oxidation of ammonia to nitrate nitrogen. Denitrification, the conversion of nitrate to N_2 and N_2O gases, occurs in persistently wet soils, resulting in a loss of soil nitrogen to the atmosphere.

Turfgrass communities growing in compacted or persistently wet soils are often invaded by various weed species. This reflects, in part, the differential adaptation of plants to poorly aerated soils. Some weed species that typically grow under these conditions may have the capacity to transmit foliar-absorbed oxygen to their roots to satisfy respiratory requirements. Thus, specific weeds may have an advantage over many turfgrasses through their ability to survive persistently wet conditions.

Temperature

Many physical, chemical, and biological events that take place in soil are strongly temperature dependent. Soil temperature is, in turn, affected by (1) atmospheric conditions (air temperature, moisture, wind, and solar radiation), (2) thermal absorption and conductivity of the soil, and (3) plant cover. Atmospheric influences on soil temperature were discussed in an earlier section of this chapter.

Thermal absorption is a function of the color, moisture level, and organic matter content of the soil. Generally, darker soils high in organic matter are more efficient in absorbing heat from the atmosphere. Heat absorption occurs faster in drier soils, since the specific heat (the amount of energy necessary to raise the temperature of 1 gram of a substance by 1°C) for water, dry mineral soil, and dry humus is 1.0, 0.2, and 0.4, respectively. Therefore, as the water content of a given soil increases, the amount of energy from solar radiation or atmospheric air required to raise its temperature increases proportionately.

Changes in soil temperature are influenced by the air-moisture-solid balance in the soil. Sandy soils warm and cool at a faster rate than clayey soils due to generally higher aeration porosity and lower retained moisture. Similarly, compacted soils undergo slower warming and cooling than well-structured soils. Temperature fluctuations during the winter months in cooler climates can be beneficial in promoting desirable soil structure. Alternate freezing and thawing of soil and wetting and drying cycles tend to rearrange compacted soil particles in such a way that aeration porosity is increased. Other temperature-dependent processes, including the formation and expansion of ice crystals and the shrinking and swelling of organic matter, also result in improved soil structure.

Most reactions within the soil occur more rapidly at higher temperatures. Microbial activity, which is so important for nitrogen transformations, organic matter decomposition, and other processes, is highly temperature dependent.

Adaptation of turfgrass species is considerably influenced by soil temperature. Root growth of Kentucky bluegrass is slowed at soil temperatures above 75°F, while 85°F is favorable to bermudagrasses.

Soil Chemistry

Soil chemistry deals with the chemical reactions that occur on colloidal surfaces in the soil. A knowledge of soil chemistry is essential for understanding nutrient availability to plants, phytotoxic effects from soil constituents, and the relationship between fertility and the physical and biological properties of soil.

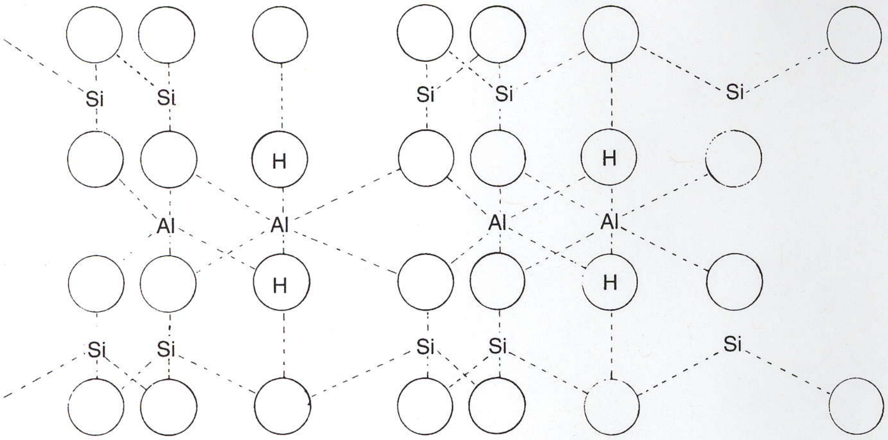


Figure 4.21. Clay particle configuration with many layers, each consisting of several planes of oxygen atoms with intervening silicon and aluminum ions (two-to-one-type clay).

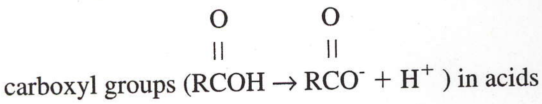
Soil Colloids

Soil colloids include small soil particles measuring 0.2 mm or less in diameter. Clay and humus make up the colloidal component of a soil. Most clay colloids are secondary crystalline minerals formed from such primary minerals as quartz, feldspars, micas, hornblende, and augite. Clay colloids are made up of planes of oxygen atoms (O, OH) with silicon (Si^{4+}) and aluminum (Al^{3+}) atoms holding the oxygens together by ionic bonding (Figure 4.21). Several planes of oxygen atoms with intervening silicon and aluminum planes make up each crystal layer within a clay particle, which has many layers stacked like a deck of cards. Silicon and aluminum atoms making up the cationic planes within the crystal layer may be substituted with cations of lower valence. For example, Al^{3+} may be substituted for Si^{4+}, Mg^{2+}, and Fe^{2+}, or Zn^{2+} for Al^{3+}. This is called isomorphous substitution and results in unsatisfied negative charges in adjacent oxygen atoms. The net effect is the colloid's negative surface charges and consequent capacity for attracting cations (*cation exchange capacity*). Other sources of negative charges include unsatisfied edge-of-clay oxygens, ionized hydrogen from hydroxyl groups, and ionized hydrogen from organic materials.

Not all clays are alike. Most have a definite repeating arrangement of component elements (*crystalline structure*). Others are not formed from well-oriented crystals and, therefore, are said to have an amorphous structure. In temperate climates the silicate clays are predominant. Sesquioxide clays, which are hydrous oxide clays of iron and aluminum (rather than silicon), are found mostly in tropical climates. Within the silicate clays, composition of the crystal layer varies depending on the number of oxygen and cationic (Si^{4+}, Al^{3+}) planes. A one-to-one-type clay has one plane of silicon and one plane of aluminum cations, along with two outer planes and a shared inner plane of oxygens. A two-to-one-type clay has two planes of silicon and one of aluminum with two outer planes and two shared planes of oxygens, as in Figure 4.21. Kaolinite is an example of a one-to-one-type silicate clay. It has almost no isomorphous substitution; its cation exchange capacity is low; and because adjacent crystal layers are strongly attracted to each other by hydrogen bonds, water does not penetrate between layers and almost no swelling of the colloids occurs. Montmorillonite is a two-to-one-type silicate clay. It has a relatively high cation exchange capacity due to considerable isomorphous substitution, and adjacent crystal layers allow water to penetrate between them. This causes substantial swelling of the clay particle and subsequent shrinking upon drying. Illite is a two-to-one-type silicate clay similar to montmorillonite except that potassium ions hold adjacent crystal layers together, thus limiting swelling.

Sesquioxide clays have a crystal layer composed of a single plane of iron and aluminum cations with planes of hydroxyl ions on either side. These clays do not swell, are not sticky, and behave more like fine sands than silicate clays.

Humus is a semistable end product of decomposing plant and animal residues. Its organic components continue to decompose, but very slowly compared to the original organic material. Humus colloids are amorphous organic particles consisting of proteins, lignins, and complex sugars. On a dry-weight basis, humus has a cation exchange capacity that is many times greater than that of clay colloids. Negative charges are due to ionization of hydrogen from oxygen-containing groups including the following:



carboxyl groups ($R\overset{\overset{O}{\|}}{C}OH \rightarrow R\overset{\overset{O}{\|}}{C}O^- + H^+$) in acids

and

hydroxyl groups ($ROH \rightarrow RO^- + H^+$) in alcohols and phenols

Cation Exchange

Positively charged ions (cations) are adsorbed onto colloids at sites with unneutralized negative charges. The adsorbing force is electrostatic attraction between the positively and negatively charged entities (Figure 4.22). The adsorbed cations resist removal by leaching water, but they can be replaced by other cations in solution due to mass action and the preferential adsorption of some cations over others. This exchange of one cation for another is called cation exchange. When such cations as NH_4^+ (ammonium), K^+ (potassium), and Ca^{2+} (calcium) are supplied to the soil by fertilizers and lime, these ions replace other cations already adsorbed to exchange sites on the surfaces of soil colloids and plant roots.

The distribution of different cations on exchange sites can dramatically affect soil physical properties. Saturation of the exchange sites by Na^+ (sodium) causes a dispersal of clay particles and a consequent loss of desirable soil structure. Sodium-saturated soils

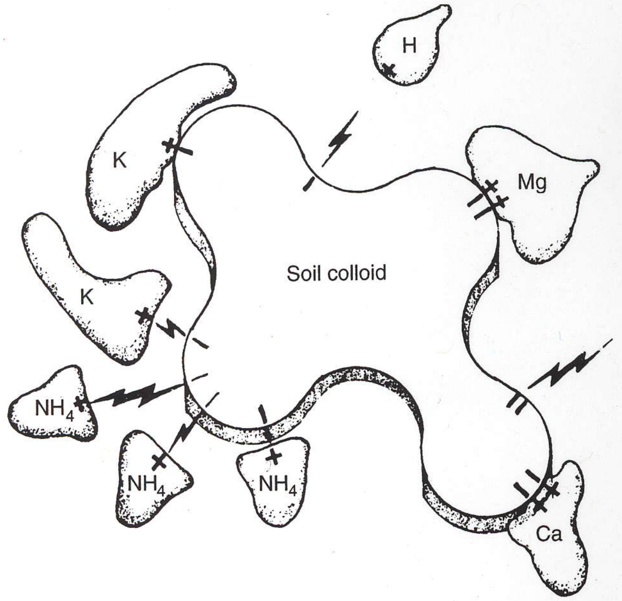


Figure 4.22. Electrostatic attraction between negatively charged soil colloid and positively charged ions.

become severely compacted and highly impermeable to water. Replacement of adsorbed sodium cations by calcium can improve flocculation of clay particles, resulting in an overall improvement in soil physical properties.

Cation exchange capacity (CEC) is the amount of exchangeable cations per unit dry weight of soil. It is measured in milliequivalents (mEq) per 100 grams (g) of soil.

The amount of a specific cation in the soil (in pounds per acre) can be calculated from CEC values. First, determine the milliequivalent weight of the cation by dividing its molecular weight by its valence. Calcium (Ca^{2+}) has a molecular weight of 40 g/mole and a valence of 2 (two positive charges). Therefore, 40 g/mole × 2 = 20 g/equivalent or 20 mg/mEq. Then, multiply the milliequivalent weight (20 mg/mEq) by the milliequivalent value for calcium given as part of the total cation exchange capacity. If the calcium portion of CEC is determined to be 13 mEq/100 g soil, then 20 mg/mEq × 13 mEq/100 g = 260 mg/100 g or 260 mg/100,000 mg or 2600 ppm (parts per million) calcium. Finally, since the surface 6 2/3 inches of an acre of field soil (acre furrow slice) weighs approximately 2 million pounds, 2600 ppm Ca is equal to 5200 lb Ca/acre. The principal cations and their equivalent weights are shown in Table 4.2.

A soil with a cation exchange capacity of 20 mEq/100 g may have 65% of the exchange sites satisfied with calcium (13 mEq/100 g), 10% with magnesium (2 mEq/100 g), 5% with potassium (1 mEq/100 g), and the rest with sodium, ammonium, hydrogen, and other miscellaneous cations. The sum of all milliequivalents of all cations adsorbed onto exchange sites in each 100 g of soil is the cation exchange capacity.

Table 4.2. Equivalent Weights of Principal Cations

Cation	Molecular Weight (g/mole)	Valence	Milliequivalent Weight (mg/mEq)
Calcium (Ca^{2+})	40	2	20
Magnesium (Mg^{2+})	24	2	12
Potassium (K^+)	39	1	39
Sodium (Na^+)	23	1	23
Ammonium (NH_4^+)	18	1	18
Hydrogen (H^+)	1	1	1

The CEC of a soil depends on its texture, type of clay, percent organic matter, and possibly pH. Fine-textured soils typically contain more clay and, therefore, have higher CECs than coarse-textured soils. The specific type of clay is important, since montmorillonite ranges from 80 to 150 mEq/100 g, while kaolinite has CECs between 3 and 15 mEq/100 g. Organic matter varies widely in CEC, depending on origin and degree of decomposition; some types contribute significantly to soil CEC due to the multiple adsorptive sites on organic colloids for retaining cations. Since many of these adsorptive sites are the result of hydrogen ionization, increasing pH usually results in more ionization and, therefore, higher CEC.

Soil Reaction

Soil reaction is an indication of the acidity or alkalinity of a soil and is measured in units of pH. Soil pH is actually the negative logarithm of the hydrogen (H^+) ion concentration. The pH scale extends from 0 to 14; each whole-number unit reflects one magnitude of change in the hydrogen (or hydroxide OH^-) ion concentration. At a pH of 7, the H^+ and OH^- concentrations are both 10^{-7} (0.0000001) moles/liter, and the soil solution is said to be at neutrality. As the pH decreases to 6, the H^+ concentration increases to 10^{-6} (0.000001). The solution becomes more acidic by a factor of 10 (one magnitude) with each pH unit decrease. At pH units above 7, the OH^- concentration is greater than the H^+ concentration, and the solution is said to be alkaline. The relationship between H^+ and OH^- ions in the soil solution is expressed by the formula

$$(H^+)(OH^-)/(HOH) = 10^{-14}$$

which states that the product of the H^+ and OH^- ion concentrations always equals 10^{-14}.

Soil pH can be measured with a pH meter, which actually measures the H^+ concentration in the soil solution. This is called active acidity. Because of the exchangeable H^+ and Al^{3+} on soil colloids, soils also have potential acidity. An equilibrium exists between active and potential acidity. If bases are added to neutralize active acidity, potential acidity is released from the exchange sites to maintain equilibrium. Thus, soils containing appreciable amounts of clay and organic matter resist pH change and are said to be buffered due to their relatively high cation exchange capacities and, therefore, high potential acidities.

The normal pH range in soil is 4.0 to 8.0. In areas where sufficient precipitation occurs to leach soluble basic salts, the soil pH tends to decrease. Irrigation may induce a similar effect, except where the irrigation water contains basic salts that accumulate in the soil. In dry regions where evapotranspiration exceeds precipitation, soil pH tends to be alkaline due to the accumulation of basic salts in the surface soil. Sodic soils (high in sodium) may reach a pH of 8.5 to 10.

Turfgrass species are adapted to a wide range of soil pHs; however, optimum growing conditions usually exist where the pH is neutral to slightly acid (7.0 to 6.0). Probably the most important detrimental effect of excessive acidity is the high solubility of aluminum and manganese, which can reach toxic concentrations in acid soils. Soil pH can be adjusted upward through periodic applications of lime, which result in better rooting, increased turfgrass vigor, improved availability of some plant nutrients, reduced availability of toxic elements, and more favorable microbial activity. Liming materials include quicklime [CaO], hydrated lime [$Ca(OH)_2$], calcitic limestone [$CaCO_3$], and dolomitic limestone [$CaCO_3$–$MgCO_3$]. Lime moves through the soil profile very slowly; therefore, applications are especially effective when made prior to establishing a new turf so that the lime can be tilled in, or possibly in conjunction with core cultivation of established turfs. The amount of lime necessary for neutralizing excessive soil acidity varies with soil texture. Clayey soils may require nearly twice as much lime as sandy soils for comparable pH adjustments. Results from soil-testing laboratories usually include recommendations for lime application where needed. Depending on the type of lime used, the amount applied at any one time should not exceed 25 to 50 lb/1000 ft^2. If hydroxide [$Ca(OH)_2$] or oxide [CaO] forms are used, rates higher than 25 pounds may severely injure the turfgrass, especially when atmospheric temperatures are high.

Excessive soil alkalinity can result in deficiencies of several plant micronutrients. The problem often can be reduced by additions of acidifying fertilizers, such as elemental sulfur and sulfates of ammonium, iron, and aluminum. As with pH adjustments by liming, the amount of sulfur or sulfates required depends on the buffering (cation exchange capacity) inherent in the soil, free $CaCO_3$ in the soil, and dissolved salts in the irrigation water. The maximum rate of finely ground elemental sulfur that should be applied at any one time is 200 lbs/acre/year on lawns and fairways, 100 lbs/acre/year on greens, and 35 lbs/acre/year on annual bluegrass greens. Due to the high potential for sulfur-induced injury, applications should be confined to periods of moderate to low atmospheric temperature.

Salted Soils

Salted soils are those containing high concentrations of soluble salts, sodium, or both. High soluble salt concentrations in the soil can seriously limit water absorption by plants, resulting in a phenomenon called *physiological drought.* The salt concentration of a soil solution is inversely related to the osmotic potential (Ψ_O), a component of soil water potential (Ψ_{SW}). As salt concentration increases, the osmotic potential and thus the soil water potential decline in value. If the soil water potential becomes as low as, or lower than, the water potential within the plant roots, plant-available water declines, regardless of the amount of water in the soil. This results in serious water deficits in the plants, the consequences of which include increased wilting, desiccation, and respiration, and reduced cell turgor, leaf size, CO_2 exchange (due to partial stomatal closure), photosynthesis, transpiration, and growth (as cell expansion requires adequate water). And because of these effects, the plants are more susceptible to environmental stresses, including wear.

Where an abundance of sodium occurs on exchange sites of clay particles, soil structure deteriorates, resulting in the following effects: macropores are destroyed and micropores dominate; pore continuity declines; water infiltration, percolation, and drainage decrease; moisture retention increases; oxygen status declines due to low oxygen diffusion; and soil strength (hardness) increases. The process by which soil permeability declines begins with an increase in the sodium percentage on exchange sites. Sodium displaces divalent cations such as calcium and magnesium. Because of the increased hydration level associated with monovalent cations, especially sodium, the thickness of the cationic layer between clay particles increases and, as a consequence, the attraction between these particles is reduced. With additional displacement, individual clay particles begin to disperse and soil structure is destroyed. Even in very sandy soils, dispersed clay particles can migrate with water movement to plug pore channels and reduce water percolation.

Soluble salts are inorganic chemicals that have a water solubility greater than 2.4 g/liter (the solubility of gypsum). The amount of total soluble salts in a soil can be estimated by measuring the electrical conductivity (EC) of the soil solution. The procedure involves mixing water with a weighed soil sample to form a saturated paste, removing the water by suction filtration, and measuring the EC of the extract (called ECe) in decisiemens per meter (dS/m). Electrical conductivity measurements can be converted to osmotic potential (Ψ_O, expressed in kilopascals [kPa]) by multiplying the ECe by –36; thus, an ECe of 1 dS/m equals –36 kPa. If the ECe is below 4 dS/m ($\Psi_O < -144$ kPa), the soil is nonsalty. The exchangeable sodium percentage (ESP) is the percentage of total exchangeable cations that is sodium.

In soils, the salinity hazard associated with increasing salinity levels is as follows: low where ECe is less than 1.5 dS/m, moderate where ECe is between 1.6 and 3.9 dS/m, high where ECe is between 4.0 and 5.0 dS/m, and very high where ECe is greater than 5.0 dS/m. The sodium hazard associated with increasing ESP levels is as follows: low where ESP is less than 3%, moderate where ESP is between 3% and 9%, high where ESP is between 9% and 15%, and very high where ESP is greater than 15%.

Salted soils are classified on the basis of two criteria: soluble salt content and exchangeable sodium percentage. The three categories of salted soils are saline, sodic, and saline-sodic. Characteristics of *saline* soils include ECe > 4 dS/m, ESP $< 15\%$ of CEC, and pH < 8.5. Saline soils are usually well structured and quite permeable. On the other hand, *sodic* soils contain sufficient Na^+ to raise the pH to 8.5 or higher, ECe is less than 4 dS/m, and ESP is 15% or more of CEC. High concentrations of Na^+ cause dispersion of clay particles, resulting in a structureless, impermeable soil. *Saline-sodic* soils are high in both sodium (ESP $> 15\%$ of CEC) and soluble salts (ECe > 4 dS/m).

Controlling problems associated with saline soils depends on reducing the concentration of soluble salts. Irrigation must be performed beyond plant moisture requirements in order to leach excess salts from the turfgrass root zone, and it must be frequent enough to prevent drying of the soil so that salts do not move upward in capillary flow. On sites where the intensity and frequency of irrigation cannot be sufficient to adequately reduce salt concentrations in the root zone, or where the irrigation water contains high concentrations of salts, turfgrass species that are tolerant of saline conditions must be used. The most salt-tolerant warm-season turfgrasses are seashore paspalum and St. Augustinegrass, while the most salt-tolerant cool-season turfgrasses are alkaligrass and the wheatgrasses (Table 4.3).

In sodic soils, exchangeable Na^+ must be replaced by other cations that do not disperse clay particles but, rather, will cause flocculation to occur. The principal cations used for this purpose include Ca^{2+} and H^+ generated from gypsum ($CaSO_4$), sulfur, or other materials. Gypsum usually causes the fastest response due to the direct exchange of Ca^{2+} for Na^+ on colloidal surfaces. This remedy is most effective when the material can be incorporated into the soil as opposed to being surface applied. The gypsum requirement (GR) is the amount of gypsum, in tons per acre, required to reclaim the soil and is calculated by the formula $GR = 1.72\ (Na_x)$, in which Na_x is the milliequivalents of sodium per 100 grams of soil (mEq/100 g) to be replaced. For example, if the exchangeable sodium has been measured to be 10 mEq/100 g (CEC = 28 mEq/100 g; therefore, ESP = 36%), and the desired exchangeable sodium is 3 mEq/100 g, then $GR = 1.72\ (10–3) = 12$ tons of gypsum per acre. Gypsum is usually surface applied at 1 ton per acre; up to 2 tons per acre per year are permissible. Elemental sulfur can be used to reduce sodium in soils with

Table 4.3. Turfgrass Salinity Tolerance

Tolerant (ECe 6.1–10)	Moderately Tolerant (ECe 3.1–6.0)	Moderately Sensitive (ECe 1.6–3.0)	Highly Sensitive (ECe <1.5)
seashore paspalum (8.6)	buffalograss (5.3)	Kentucky bluegrass (3.0)	centipedegrass (1.5)
alkaligrass (8.5)	blue grama (5.2)	zoysiagrass (2.4)	annual bluegrass (1.5) colonial bentgrass (1.5)
wheat grasses (8.0)	hard fescue (4.5)		
kikuyugrass (8.0)	strong creeping red fescue (4.5)		rough bluegrass (1.5)
St. Augustinegrass (6.5)	common bermudagrass (4.3)		
tall fescue (6.5)	hybrid bermudagrass (3.7)		
perennial ryegrass (6.3) slender creeping red fescue (6.3)	creeping bentgrass (3.7)		

free lime present, while gypsum can be used in soils with or without free lime present. The efficiency of sulfur in reclaiming sodic soils is approximately 5.6 times that of gypsum; therefore, only about 2 tons of sulfur per acre would be required to achieve the same effect. However, since sulfur must be oxidized microbially to sulfuric acid (H_2SO_4)before it can yield H^+ ions to replace exchangeable Na^+, reaction time would be much slower than with gypsum, which directly supplies Ca^{2+}.

Soil Biomass

Soil biomass consists of the sum total of all living organisms and their residues that become part of the soil profile. *Live biomass* is that portion of soil biomass that includes an array of microscopic and macroscopic organisms, plant roots, and subsurface shoots. *Residual biomass,* or dead organic matter, is made up of undecomposed and partially decomposed remnants of plant and animal material occurring within and above the soil. Given the dynamic nature of soil biomass, changes within and above the soil dramatically influence the suitability of the medium for sustaining turfgrasses.

Live Biomass

The soil is virtually alive with huge populations of microscopic plants called *microflora* (bacteria, fungi, actinomycetes, algae) and microscopic animals called *microfauna* (protozoa, nematodes). Occasionally, we find evidence of larger, soil-inhabiting animals such as earthworms, arthropods (insects, mites, centipedes, millipedes), gastropods (slugs, snails), and even larger burrowing animals including moles, gophers, and mice. These are referred to collectively as *macrofauna.* Live roots and other plant parts constitute the soil *macroflora.*

Soil organisms are largely beneficial in that they decompose organic materials, convert materials into plant-available forms, and improve soil structure. Some, however, are pests that can cause extensive damage to turf.

Of the soil microflora, bacteria are the most numerous. Bacteria occur in two principal categories: *autotrophic* bacteria, which derive their nutritive carbon from CO_2, and *heterotrophic* bacteria, which obtain carbon from organic matter. Autotrophs are very beneficial because of the processes by which they obtain energy. Oxidation of ammonia to nitrites and then to nitrates, as well as oxidation of sulfur, iron, manganese, and other elements, is an important process for providing plant-available nutrients. Conversion of carbon monoxide (CO) to methane (CH_4), a reduction reaction, or to CO_2, an oxidation reaction, is certainly important in highly populated areas where CO emissions would otherwise accumulate to phytotoxic concentrations in the atmosphere. Some autotrophs engage in less desirable activities, such as denitrification, or the reduction of nitrate (NO_3^-) to N_2 or N_2O gases. This may account for substantial losses of fertilizer nitrogen to the atmosphere, especially from warm persistently wet soils.

Heterotrophs include most of the soil bacteria. Soils may contain more than 100 million bacteria per gram of dry soil. The optimum pH for most bacteria is near neutrality.

Bacteria are the principal organisms for decomposing organic matter and, as such, provide for recycling of nutrients contained within plant and animal residues. Were it not for heterotrophic bacteria, turfgrass and other plant communities (indeed, all organisms) would be virtually inundated by their own debris. Some heterotrophs fix atmospheric nitrogen and, thus, enable plants to obtain nitrogen from the air. Certain nitrogen-fixing heterotrophs attack root hairs of host plants, resulting in the formation of protective nodules. A symbiotic union is thus formed in which the plant supplies organic substances and minerals to the bacteria, and the bacteria fix and supply nitrogen to their host plant. Fixation actually occurs as the bacteria utilize atmospheric nitrogen for synthesizing body proteins. Since the life span of a single bacterium is only a few hours, a portion of the bacterial population is continually dying, decomposing, and releasing ammonium and nitrate ions to host and adjacent plants. The principal turf-type host plants that support nitrogen-fixing bacteria are the clovers.

Compared with other microflora, fungi are present in the lowest numbers in soil, especially sands. Usually, their numbers range from 1,000 to 1 million per gram of dry soil. They are most competitive at a relatively low soil pH. Fungi occur in three primary categories depending on their nutritive processes: parasitic, saprophytic, and symbiotic fungi. Parasitic fungi that cause plant diseases will be discussed in Chapter 7. Saprophytes are fungi that function as decomposers of residual biomass. Some are especially important for breaking down cellulose and lignin, which are resistant to many other types of biodecomposition. Symbiotic fungi form an intimate association with some plant roots that aids in the absorption of specific nutrients. This association is referred to as *mycorrhizae.*

Actinomycetes tend to dominate in high-pH soils. Usually, they are second to bacteria in numbers per gram of dry soil. They also give moist soil its characteristic smell. Actinomycetes are widely known for their production of antibiotics. They are important in turf soils as decomposers of residual biomass, especially cellulose and other resistant forms.

Algae are chlorophyll-containing microflora that serve as a source of organic matter. Some types function in nitrogen fixation. In turf, algal mats sometimes develop following severe disease incidence. Upon drying, they form a hard crust that, unless removed or broken up, makes the surface nearly impervious to water.

The microfauna include protozoa and nematodes. Protozoa are primitive, single-celled animals that feed mainly on bacteria. Nematodes are threadlike worms that are widely distributed in agricultural soils. Predaceous nematodes prey on soil flora and fauna, including other nematodes. Parasitic nematodes are of greatest concern because of their ability to infest plant roots and severely damage turfgrasses. Their entrance into plants facilitates entry by other pathogens. Control of turfgrass-parasitic nematodes will be discussed in Chapter 7.

Prominent among macrofauna are the earthworms. Of the 7000 species that have been identified, three are most commonly encountered. These are the garden worm (*Helodrilus calinginosus*), the red worm (*Helodrilus foetidus*), and the night crawler (*Lumbricus terrestris*). Earthworms do not feed on live plants but are extremely effective in reducing accumulations of residual organic materials. The ingested organic material and soil are excreted as small, granular aggregates containing substantial amounts of plant nutrients. The deposited casts are sometimes objectionable in closely mowed turf;

however, the beneficial effects of earthworms, which include the creation of numerous channels (called *biogenic megapores*) through much of the soil, often outweigh any problems resulting from their activity. In studies conducted in Illinois, control of earthworm populations by some pesticides resulted in thatch development in an otherwise thatch-free Kentucky bluegrass turf. Other effects included higher soil bulk densities, reduced water infiltration and hydraulic conductivity, lower soil organic matter concentrations, shallow rooting, higher wilting proneness, and greater disease severity. All of the pesticides used in these studies (e.g., calcium arsenate, bandane, chlordane) are no longer commercially available. Earthworms are suppressed by applications of thiophenate-methyl and bendiocarb; however, these pesticides are not registered for this use. Although present throughout the growing season, the earthworms appeared to be more active at the surface during spring and fall.

Occasionally, masses of dead earthworms may be present at the turf surface. This usually occurs when the sun suddenly appears following rainfall. At other times the surface of the ground may be disrupted by earthworm activity. Under these circumstances, it is tempting to schedule a pesticide application for control. However, the long-term advantages and disadvantages should be carefully assessed before deciding that a chemical control measure will invariably result in better turf.

Several arthropods and gastropods serve as decomposers of organic debris. Others, particularly certain insects, cause serious damage to turf. The appearance of a few of these small animals does not necessarily dictate that some control measure should be undertaken. Many insects or other small animals occurring in turf do not cause injury, and some of the potentially injurious ones are of no importance unless their populations build up to high levels. Strategies for controlling insects and other pests will be discussed further in Chapter 7.

Depending on location, numerous larger animals may occasionally be troublesome because of their burrowing into the turf. As with other macrofauna, their activities may be beneficial. However, when serious damage occurs or is threatened, control measures may be appropriately undertaken (see Chapter 7).

The soil macroflora include turfgrasses and other plants that are anchored to the soil. Root, rhizome, and basal shoot growth are influenced by soil conditions and, in turn, influence the physical, chemical, and biological properties of the soil. Where a fairly contiguous plant community (turfgrass) covers the soil there is a moderation of soil temperature and moisture extremes. The protective cover provided by turfgrasses reduces soil compaction from traffic, splashing raindrops, and other physical forces. Growth of roots and other belowground plant organs stirs the soil and promotes better structure. Deposition of plant debris within and atop the soil provides residual biomass, which is the carbon and energy source for many beneficial soil organisms. Decomposition of organic residues brings about numerous benefits measurable as improved soil properties and plant growth.

Residual Biomass

Residual biomass includes the undecomposed and partially decomposed organic residues that occur within the soil and above the soil surface. Cultivated mineral soils usually contain

1% to 5% organic matter. Turf soils may have somewhat higher percentages, especially near the surface. Three distinct types of residual biomass are common in turf: thatch, mat, and the organic fraction usually associated with cultivated soils.

Thatch is the layer of organic residue located immediately above the soil surface. It may be largely undecomposed or at an advanced stage of decomposition, especially where it meets the soil. Although considerable attention has been devoted to processes of thatch formation, its development in turf is still not completely understood. A typical explanation is that thatch occurs when plant biomass production exceeds decomposition, resulting in an accumulation of surface debris followed by growth of the turfgrass plants within this debris. Obviously, factors that suppress the decomposition rate or promote excessive production could trigger this imbalance. Another possible explanation is that growth of adventitious roots and stolons above the soil surface provides a medium within which dead or dying plant residues accumulate and decompose more slowly than in soil. Regardless of the mechanism of formation, the living plant community apparently exerts a stabilizing influence on thatch, since it has been shown that death of the turfgrass community is often followed by destruction of the thatch layer. Limited work to characterize thatch as a turfgrass growth medium has shown that it is a well-aerated, compaction-resistant medium with poor nutrient- and water-holding properties. Therefore, turfgrasses with most of their roots confined to the thatch layer are more prone to injury from environmental stresses. Diseases and insect injury are generally more severe where thatch is a problem. Measures for controlling thatch will be discussed in Chapter 6.

Mat is a surface layer of organic residue mixed with soil. Often mat results where soil has been incorporated into a thatch layer. Mechanisms by which this can occur include some naturally occurring phenomena, such as earthworms depositing soil within the thatch or wind depositing soil onto thatchy turf, and some culturally induced transformations, such as topdressing with soil or reincorporation of soil following core cultivation. Mat is generally considered a desirable feature in turf, especially in those turfs receiving frequent or intensive traffic. A mat layer not only provides better surface resiliency but seems to stabilize turf against impacting forces encountered in athletic activities. Although a mat layer may not be easily seen in a turf profile, its presence can be qualitatively determined by carefully working a soil core (taken with a soil probe) with the fingers, beginning at the base of the core. At moderate moisture, the soil should be easily separated from the roots until the mat layer is reached; then, the soil resists separation due to the organic material interspersed throughout the surface soil.

In addition to thatch and mat, other types of organic residues occur in mineral soils. These result from incorporating organic amendments during establishment or from deposition of organic materials during soil formation and under cropping practices. Since the soil organic matter is constantly undergoing change, it must be replenished continuously to maintain soil productivity. The principal replenishing sources are plant roots, residues of other plant parts, and soil microorganisms. Decomposition of organic matter results in the use of some carbon, nitrogen, and other elements by microorganisms; release of carbon compounds, water, and other elements to the soil and to the atmosphere; and the formation of a semistable organic residue called *humus*.

Nitrogen is the nutrient that most often controls the rate of organic matter decomposition because it is used to synthesize proteins in new microbial populations. The ratio of carbon to nitrogen, called the C:N ratio, controls the decomposition rate and largely determines whether nitrogen will be released to the soil or pulled from it by microbial activity. Most organic materials have a carbon content of about 40%; however, the nitrogen content is variable. Bacteria have C:N ratios of 4:1 to 5:1, while fungi have a C:N ratio of 9:1. Plant residues having C:N ratios of 20:1 or narrower contain sufficient nitrogen to supply microorganisms and also release nitrogen for plant use. This is because in the decomposition of organic matter much carbon is released to the atmosphere as CO_2, while most of the nitrogen is reused repeatedly by microbial populations, thus narrowing the C:N ratio. As the organic matter is used up, microorganisms die, decompose, and release more CO_2 as well as plant-available nitrogen. When organic materials with large C:N ratios, such as sawdust (250:1) and straw (90:1), are added to the soil, nitrogen is pulled from adjacent soil. Since microorganisms can compete more effectively than plant roots for available nitrogen within the soil, plants may show symptoms of nitrogen deficiency unless additional nitrogen is supplied through fertilization.

The organic matter of greatest importance in soil is that which is undergoing rapid decomposition. This is called the *active organic matter.* The principal benefits derived from active organic matter include soil aggregation and stabilization of soil aggregates by organic gums that cement clay particles together; mineralization (release) of plant nutrients; and production of humus, which provides additional cation exchange capacity. Decomposition is favored by conditions that favor microbial activity, including adequate aeration and moisture, slightly acid to neutral pH, available nutrients, suitable temperatures (optimum is 95°F), and a favorable C:N ratio.

THE BIOTIC ENVIRONMENT

The biotic component of the turfgrass environment includes, primarily, the use and culture of the turf by people. Other biota are sometimes included but, except for humans, most of these have been discussed under Soil Biomass. Human activities frequently constitute the most formidable of adverse influences against turfgrass persistence and quality. Conditions that reflect the type and intensity of these activities are wear and breakage of the turf, and displacement and compaction of the soil.

Wear is the physical deterioration of a turfgrass community resulting from excessive traffic. What constitutes "excessive" traffic will vary depending on turfgrass genotype, the physiological condition of the plants, and edaphic and climatic conditions. A rough bluegrass turf in a wet, shaded environment will be much more susceptible to wear than a well-cared-for bermudagrass turf on a sunny, sandy site. However, since bermudagrass would not survive where rough bluegrass is well adapted as a permanent turfgrass, the comparison is somewhat misleading. Additional comparisons of turfgrass species for wear tolerance are given in Chapter 3.

Direct pressure applied to the turf tends to crush shoots, especially when the plants are stiff with frost or wilted. When the pressure is accompanied by lateral forces, wear is greater. Since wear tolerance varies so much from site to site, and at different times on the same site, a traffic threshold level must be determined for each location and time of year. Then measures can be taken to control traffic intensity so that the threshold level is not exceeded. In many situations, wear can be minimized by making some simple adjustments. On golf courses, traffic patterns on greens, approaches, exit lanes, and tees can be made less concentrated by coordinating placement of pins and tee markers. Shifting sandtrap positions to provide wider traffic lanes and improving drainage can dramatically reduce wear. On athletic fields used for practice, intensive traffic can be shifted to different field locations to reduce wear and promote recovery of worn turf. Strategic placement of ornamental plantings, sidewalks, fences, and other landscape features is often effective in uniformly distributing or concentrating traffic for the purpose of promoting better turf. With a little imagination, small provisions that will result in less wear can be made on almost any site.

In succeeding chapters cultural practices are discussed in detail. Much of this discussion is oriented toward achieving healthy, vigorous turfgrass that is highly tolerant of environmental stresses, including traffic. Some rather basic cultural considerations for improving wearability include raising the mowing height to provide larger, more vigorous shoots; reducing plant succulence by adjusting fertilization and irrigation practices; and cultivating to promote better shoot and root growth. Some conditions require that traffic be avoided entirely. Frosted leaves, winter-dormant turfgrass, and wet soil conditions predispose the turf to serious injury from even light traffic. Moderately shaded turfgrass may be so weak that it is intolerant of all but minimal traffic.

Turf breakage results from those activities that disrupt the structural integrity of the turf. A golf ball landing on a green can tear the turf's surface, especially where plants are shallow rooted, resulting in "ball marks." If the damage is not repaired by carefully reshaping the turf's surface, mowing can scalp the upraised plants and soil from the turf causing even more severe injury. Golf shots, as well as a variety of other physical forces, may rip out sections of sod, causing "divots" in the turf. The size of the divots is influenced by the specific nature of the physical force, turfgrass species, cultural practices, and the overall condition of the turf. Repair operations may include replacing torn sections of sod or filling in the divots with soil (and perhaps seed) to promote recovery.

Soil displacement and associated rutting occurs where foot or vehicular traffic substantially changes the shape of the turf's surface. The susceptibility of a turf to soil displacement and rutting is a function of soil moisture, texture, and strength; the extent to which turfgrass roots and lateral shoots enhance the shear strength of the turf; and the intensity and duration of the traffic event. As soil displacement and rutting are most likely to occur on wet, fine-textured soils, traffic should be carefully controlled whenever and wherever these conditions exist.

Compaction is a physical condition of soil resulting from the compression of soil constituents into a relatively dense mass. A moderate amount of compaction may be desirable in turf soils to provide a firm surface and favorable contact between plant roots and soil particles. Severe compaction, however, is usually associated with inadequate soil aeration and poor drainage due to the loss of aeration porosity. Soil compaction can occur in the surface layer or at some depth within the soil profile. Subsurface compaction layers are called *pans;* the two principal types are genetic pans and induced pans. Genetic pans occur naturally due to very high clay content present during soil formation (claypan) or cementation of soil particles by organic matter or inorganic materials (hardpan). Induced pans result from pressure applied by cultivation or tillage operations. In field crop culture, a *plow sole* can develop at plow depth due to the alteration of soil physical structure from horizontal operation of the plow. With core cultivation of turf, induced pans can occur at the maximum depth of penetration of the tines and, possibly, along the sides of the holes resulting from cultivation.

A problem that is analogous to surface and subsurface compaction layers, but which occurs primarily in sandy soil profiles, is called "black layer." It is especially common where a sand topdressing program has been employed for turfs growing on finer sands or other fine-textured media. It appears as a coal black band ranging in thickness from a few millimeters to several inches. The characteristic foul-smelling odor of the band is due to the production of sulfide gases by bacteria under anaerobic conditions. The band itself is actually a precipitate of metal sulfides, including FeS, MnS, and MgS, that coats soil particles and clogs pore spaces. Some investigators believe that the existence of black layer in turf soils reflects the use of sulfur or sulfur derivatives for reducing the pH of calcareous sands. While the use of sulfur-containing materials may contribute to the formation of black layer, it is possible that the sulfur released from the biodegradation of soil organic matter may be sufficient when other conditions are favorable.

Surface compaction of turf soils occurs primarily in response to traffic. Particles of soil are pressed together with resultant increases in bulk density, soil strength, and water-holding capacity, and decreases in aeration porosity, water infiltration, and percolation. Increased bulk density results from the reduction in soil volume following compaction. The change in volume is attributed to a loss in total porosity; some pores are lost or reduced in size as clay particles become flattened in a plane perpendicular to the compacting force. The large (aeration) pores that are so important in drainage and soil aeration are most susceptible to loss during compaction. The accompanying shift in pore-size distribution (to generally smaller pores) accounts for increased moisture retention. Capillary pores assume a higher proportion of total porosity as aeration porosity is reduced. Increased soil strength, measurable as resistance to penetration, partially accounts for the poor root growth commonly observed in compacted soils. Inadequate oxygen concentration and relatively high concentrations of CO_2 and other potentially toxic gases are additional reasons for poor rooting in compacted soils. Low water infiltration and percolation rates typically accompany shifts in pore-size distribution toward smaller pores.

Physiological responses of plants to soil compaction include not only reduced root growth but reduced shoot growth, reduced water and nutrient uptake, reduced tolerance to heat and drought stresses, and increased susceptibility to scald and direct low-temperature injury. The generally weakened condition of plants growing in compacted soils coupled with prolonged periods of high humidity in the plants' microenvironment account for higher disease incidence as well.

Ecological responses of turfgrass communities subjected to soil compaction are frequently observed as dramatic shifts to compaction-tolerant weed populations such as knotweed, annual bluegrass, goosegrass, and white clover.

Soil compaction influences turfgrass cultural requirements. More frequent fertilization and irrigation are required to compensate for restricted rooting and generally poor growth. Cultivation practices are necessary to improve infiltration, drainage, and plant growth. Installation of drainage tiles, slit trenches, and catch basins may be essential to dispose of excessive surface moisture. Pesticides are required to control the higher incidence of weeds and diseases that frequently accompany soil compaction. Furthermore, mistakes in the application of pesticides that result in damage to the turf are of greater consequence where, because of compaction, suitable conditions for turfgrass recovery do not exist.

Methods for preventing soil compaction or reducing its effects will be discussed further in Chapters 6 and 8.

QUESTIONS

1. Differentiate among reflection, absorption, transmission, and conversion (to chemical energy through photosynthesis) of light by a turfgrass plant.
2. How is a turfgrass plant likely to respond to reduced light intensity?
3. How should cultural practices be adjusted to sustain a turfgrass community under low light intensities?
4. Explain the processes by which turfgrasses dissipate heat energy.
5. What factors influence turfgrass aerial shoot and root temperatures?
6. Explain the significance of moisture in turfgrass growth and survival.
7. What influence is wind likely to have on turfgrass growth and survival?
8. What is turfgrass edaphology?
9. Characterize a block of soil (supporting the growth of turfgrass plants) with respect to the distribution of solids, water, and air.
10. List the different forms of organic residues that are likely to be found in association with turf and explain their significance in turfgrass growth and quality.
11. Differentiate between soil texture and structure.
12. Characterize the different types of soil moisture that occur in a saturated soil and explain their significance with respect to turfgrass growth and quality.
13. What are soil colloids and in what ways are they likely to contribute to turfgrass nutrition?
14. What is meant by soil reaction and what is its significance with respect to turfgrass growth?
15. Characterize the biotic environment influencing a turfgrass community.