GLOBAL EDITION

The Nature and Properties of Soils

FIFTEENTH EDITION

Ray R. Weil • Nyle C. Brady

Pearson

THE NATURE AND PROPERTIES OF SOILS

upper layers of permafrost to thaw. In parts of Alaska, for example, temperatures in top layers of permafrost have risen about 3.5 °C since the late 1980s, resulting in melting rates of about a meter in a decade. Such melting has serious implications since it can drastically affect the physical foundation of buildings and roads, as well as the stability of root zones of forests and other such vegetation in the region. Trees fall and buildings collapse as the frozen layers melt. The thawing of arctic permafrost is expected to further accelerate global warming, as decomposition of organic materials long trapped in the frozen layers of Histels releases vast quantities of carbon dioxide into the atmosphere (see Figure 3.19).

Soil with ice lenses may contain much more water than would be needed to saturate the soil in the unfrozen state. When the ice lenses thaw, the soil becomes supersaturated because the excess water cannot drain away through the underlying still-frozen soil. Soil in this condition readily turns into noncohesive mud that is very susceptible to erosion and movement by mudslides.

Soil Heating by Fire

Fire is one of the most far-reaching ecosystem disturbances in nature. In addition to the obvious aboveground effects of forest, range, or crop-residue fires, the brief but sometimes dramatic changes in soil temperature also may have lasting impacts below ground. Unless the fire is artificially stoked with added fuel, the temperature rise from wildfires is usually very brief and is limited to the upper few centimeters of soil. But the temperatures resulting from "slash and burn" practices in the tropics (see Section 20.7) may be sufficiently high in the upper few mm of soil to cause the loss of nearly all organic natter and even the breakdown of minerals such as gibbsite and kaolinite.

The heat from wildfires may also cause the breakdown and movement of organic compounds, especially in sandy soils (Figure 7.33). The high temperatures (>125 °C being common) essentially distill certain fractions of the organic matter, with some of the volatilized hydrocarbon compounds moving quickly through the soil pores to deeper, cooler areas. As these compounds reach cooler soil particles deeper in the soil, they condense (solidify) on the surface of the soil particles and fill some of the surrounding pore spaces. Some of the condensed compounds are water-repellent (hydrophobic) hydrocarbons. Consequently, when rain comes, water infiltration in such burned over sandy soil may be greatly reduced in comparison to unburned areas. This effect of soil temperature is quite common on pine forest and chaparral lands in semiarid regions and may be responsible for the disastrous mudslides that occur when the layer of soil above the hydrophobic zone becomes saturated with rainwater (see Figure 7.33*d*).

Fires also affect the germination of certain seeds, which have hard coatings that prevent them from germinating until they are heated above 70–80 °C. The burning of straw in wheat fields generates similar soil temperatures, but with the effect of killing most of the weed seeds near the surface and thus greatly reducing subsequent weed infestation. The heat and ash from low intensity fires may also hasten the cycling of plant nutrients. In fact, grassland fires often stimulate plant regrowth so much that soil organic matter increases with fire as the enhanced plant biomass more than compensates for the loss of soil organic matter during the fire. On the other hand, fires set to clear land of timber slash may burn long and hot enough to seriously deplete soil organic matter and kill so many soil organisms that forest regrowth is inhibited.

Contaminant Removal

The removal of certain organic pollutants from contaminated soils can be accomplished by raising the soil temperature, causing the offending compounds to volatilize and diffuse out of the soil in the gaseous state. The process may be prohibitively expensive if the soil has to be excavated and hauled to and from a heated extraction facility. Techniques are under development to warm the soil in place in the field using electromagnetic radiation. The resulting temperatures are sufficiently high to vaporize some contaminants, such as toxic components in diesel fuel, which can then be flushed from the soil by a stream of injected air.

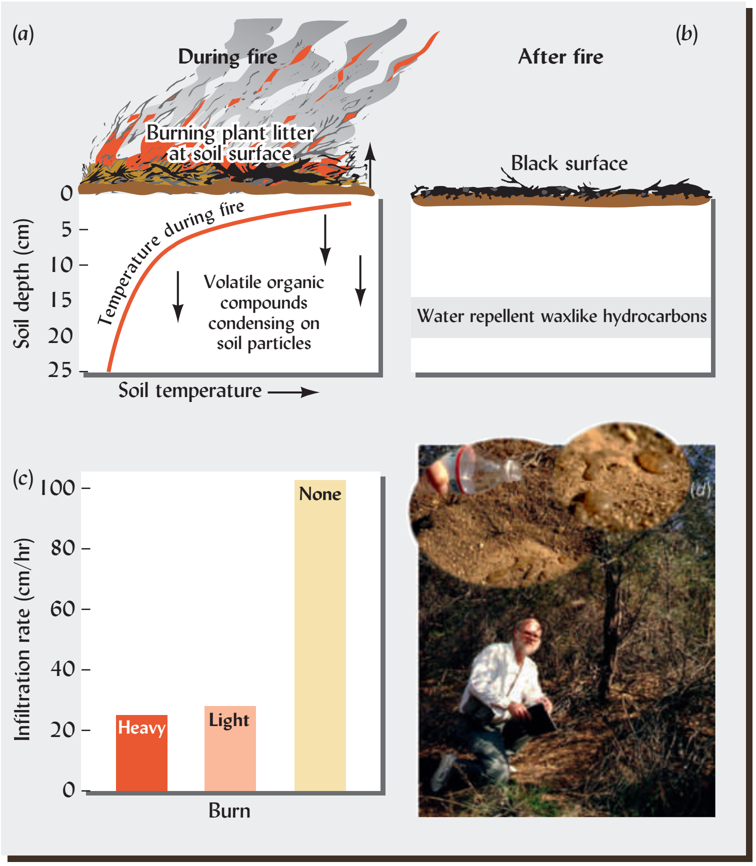

Figure 7.33 (a) *Wildfires heat up the surface layers of a sandy soil sufficiently to volatilize organic compounds from the soil organic matter and surface litter.* (b) *Some of the volatilized organics then diffuse away from the heat down into the soil and condense (solidify) on the surface of cooler soil particles. These condensed compounds are waxlike hydrocarbons that are water repellent and drastically reduce the infiltration of water into the soil for a period of years.* (d) *An ecologist investigates the soil under chaparral vegetation in southern California, USA, where dry season fires have created a water repellant layer a few cm below the surface. The inset photos show water beading up on this soil layer exposed by scrapping aside the upper 8 cm of loose, nonrepellent soil.* [Concepts from DeBano (2000), bar graph from data in Dryness (1976); photos courtesy of Ray R. Weil]

7.9 ABSORPTION AND LOSS OF SOLAR ENERGY[9]

The temperature of soils in the field is directly or indirectly dependent on at least three factors: (1) the net amount of heat energy the soil absorbs; (2) the heat energy required to bring about a given change in the temperature of a soil; and (3) the energy required for processes such as evaporation, which are constantly occurring at or near the surface of soils.

Solar radiation is the primary source of energy to heat soils. But clouds and dust particles intercept the sun's rays and absorb, scatter, or reflect most of the energy (Figure 7.34). Only about 35%–40% of the solar radiation actually reaches the Earth in cloudy humid regions and 75% in cloud-free arid areas. The global average is about 50%.

Little of the solar energy reaching the Earth actually results in soil warming. The energy is used primarily to evaporate water from the soil or leaf surfaces or is radiated or reflected back to the sky. Only about 10% is absorbed by and warms the soil. Even so, this warming is critical to soil processes and to plants growing in soils.

Albedo. The fraction of incident radiation that is reflected by the land surface is termed the **albedo** and ranges from as low as 0.1–0.2 for dark-colored, rough soil surfaces to as high as 0.5 or more for smooth, light-colored surfaces. Vegetation may affect the surface albedo either way, depending on whether it is dark green and growing or yellow and dormant.

The fact that dark-colored soils absorb more energy than lighter-colored ones does not necessarily imply, however, that dark soils are always warmer. In fact, the opposite is often true. In most landscapes, the darkest soils are those found in the low spots where excessive wetness has caused organic matter to accumulate. Therefore, the darkest soils are also usually

[9]For application of these principles to the role of soil moisture in models of global warming, see Lin et al. (2003).

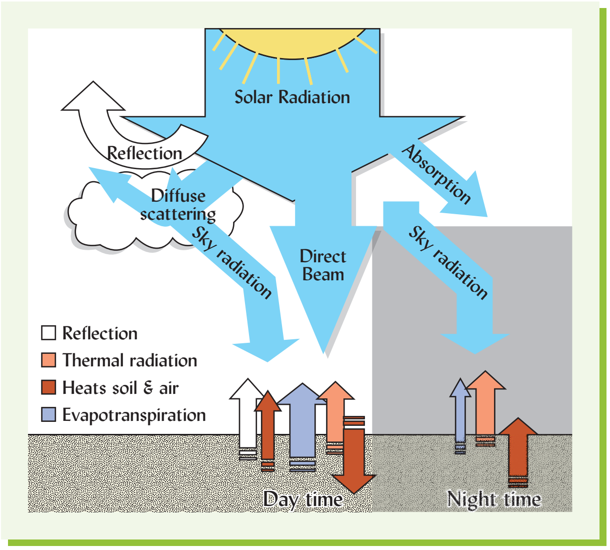

Figure 7.34 *Schematic representation of the radiation balance in daytime and nighttime in the spring or early summer in a temperate region. About half the solar radiation reaches the Earth, either directly or indirectly, from sky radiation. Most radiation energy that strikes the Earth in the daytime is used to drive evapotranspiration or is radiated back to the atmosphere. Only a small portion, perhaps 10%, actually heats the soil. At night the soil loses some heat, and some evaporation and thermal radiation occur.* (Diagram courtesy of Ray R. Weil)

the wettest. The water in these soils will absorb much energy with only little change in temperature (see Section 7.10), and it also cools the soil when it evaporates.

Slope Angle and Aspect. The angle at which the sun's rays strike the soil also influences soil temperature. If the incoming path of the rays of solar energy is perpendicular to the soil surface, energy absorption (and soil temperature increase) is greatest (Figure 7.35). The effect of the direction of slope, or **aspect**, on forest species is illustrated in the photo in Figure 7.35.

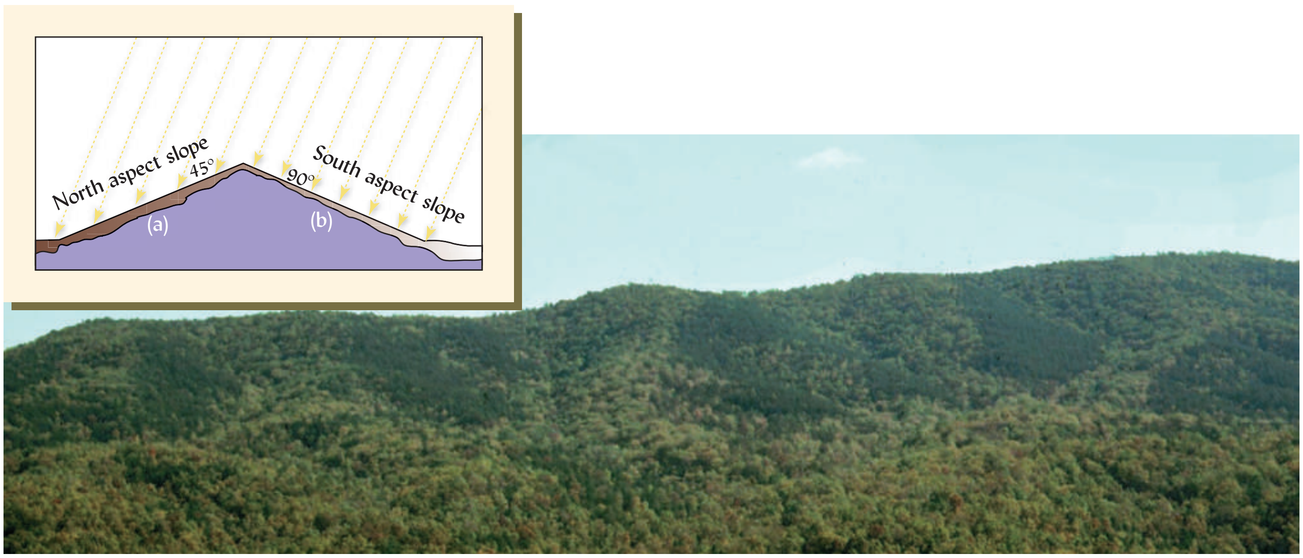

Figure 7.35 (Inset diagram) *Effect of slope aspect on solar radiation received per unit land area. Slope* (a) *is north facing and receives solar radiation at an angle of 45° to the ground surface so only 5 units of solar radiation (arrows) hit the unit of land area. The same land area on the south-facing slope* (b) *receives 7 units of radiation at a 90° angle to the ground. In other words, if a given amount of radiation from the sun strikes the soil at right angles, the radiation is concentrated in a relatively small area, and the soil warms quite rapidly. This is one of the reasons why north slopes tend to have cooler soils than south slopes. It also accounts for the colder soils in winter than in summer.* (Photo) *A view looking eastward toward a forested mountain in Virginia, USA, illustrates the temperature effect. The main ridge* (left to right) *runs north-south and the smaller side ridges run east-west (up and down). The dark patches are pine trees in this predominantly hardwood deciduous forest. The pines dominate the southern slopes on each east-west ridge. The soils on the southern slopes are warmer and therefore drier, less deeply weathered, and lower in organic matter.* (Photo and diagram courtesy of Ray R. Weil)

Planting crops on soil ridges is one method of controlling the soil aspect on a microscale. This is most effectively done at high latitudes by planting crops on the south- or southwest-facing sides of ridges. The ridges need to be only about 25 cm tall to have a major effect. For example, in Fairbanks, Alaska, midafternoon soil temperatures (at 1 cm depth) in early May on the south side of such a ridge can be about 15 °C warmer than on the north side and about 8 °C warmer than on level ground.

Rain. Mention should also be made of the effect of rain or irrigation water on soil temperature. In the summer, rainfall cools the soil, since it is often cooler than the soil it penetrates. On the other hand, in temperate zones, spring rain definitely warms the surface soil in the short term as the relatively warm water moves into cold soil. However, spring rain, by increasing the amount of solar energy used in evaporating water from the soil, can also accentuate low temperatures.

Soil Cover. Whether the soil is bare or is covered with vegetation, mulch, or snow is another factor markedly influencing the amount of solar radiation reaching the soil. Bare soils warm up more quickly and cool off more rapidly than those covered with vegetation, with snow, or with plastic mulches. Frost penetration during the winter is considerably greater in bare, noninsulated land.

Even low-growing vegetation such as turfgrass has a very noticeable influence on soil temperature and on the temperature of the surroundings (Table 7.5). Much of the cooling effect is due to heat dissipated by transpiration of water. To experience this effect, on a blistering hot day, try having a picnic on an asphalt parking lot instead of on a growing green lawn!

The effect of a dense forest is universally recognized. Timber-harvest practices that leave sufficient canopy to provide about 50% shade will likely prevent undue soil warming that could hasten the loss of soil organic matter or the onset of anaerobic conditions in wet soils. The effect of timber harvest to soil temperature as deep as 50 cm is seen in the case presented in Table 7.6, where tree removal warmed the soil in spring, even though it also raised the soil water content, since trees were no longer taking up water. However, as we shall see in the next section, a higher water content normally slows the warming of soils in spring.

7.10 THERMAL PROPERTIES OF SOILS

Specific Heat of Soils

A dry soil is more easily heated than a wet one. This is because the amount of energy required to raise the temperature of water by 1 °C (its heat capacity) is much higher than that required to warm soil solids by 1 °C. When heat capacity is expressed per unit mass—for

Table 7.5
Maximum Surface Temperatures for Four Types of Surfaces on a Sunny August Day in College Station in Texas, USA

	Maximum temperature, °C	
Type of surface	**Day**	**Night**
Green, growing turfgrass	31	24
Dry, bare soil	39	26
Brown, summer-dormant grass	52	27
Dry synthetic sports turf	70	29

Data from Beard and Green (1994).

Table 7.6

TREE REMOVAL EFFECT ON SOIL AERATION AND TEMPERATURE IN A SUBTROPICAL PINE FOREST

Removal of 55-year-old loblolly pine trees reduced transpiration and shading, resulting in a higher water table, warmer spring temperatures, and lower redox potentials in this Vertic Ochraqualf. Because warmer temperatures stimulated microbial activity, E_h was lower in spring, though the soil was wetter in winter.

		Soil temperature, °C		Soil redox potentials E_h, V	
Site treatment	Time soil is saturated, %	Winter	Spring	Winter	Spring
Measured at 50-cm depth					
Undisturbed pine stand	39	11.8	18.3	0.83	0.65
Trees cut, soil not compacted	71	11.7	20.5	0.51	0.11
Measured at 100-cm depth					
Undisturbed pine stand	57	13.3	17.3	0.83	0.49
Trees cut, soil not compacted	69	13.2	18.7	0.54	0.22

Data from Tiarks et al. (1996).

example, in calories per gram (cal/g)—it is called **specific heat** or heat capacity *c*. The specific heat of pure water is defined as 1.00 cal/g (or 4.18 joules per gram, J/g); that of dry soil is about 0.2 cal/g (0.8 J/g).

The specific heat largely controls the degree to which soils warm up in the spring, wetter soils warming more slowly than drier ones (see Box 7.4). Furthermore, if the water does not drain freely from the wet soil, it will be evaporated, a process that is very energy consuming, as the next section will show.

The high specific heat of soils is also used in the design of energy-efficient geothermic temperature control systems that both warm and cool buildings. To maximize heat-exchange contact with the soil, a network of pipes is laid underground near the building to be heated and cooled. Advantage is taken of the fact that subsoils are generally warmer than the atmosphere in the winter and cooler than the atmosphere in the summer. Fluid circulating through the network of pipes absorbs heat from the soil during the winter and releases it to the soil in the summer. The high specific heat of soils combined with their enormous mass permits a large exchange of energy to take place without greatly modifying the soil temperature.

Heat of Vaporization

The evaporation of water from soil surfaces requires a large amount of energy, 540 kilocalories (kcal) or 2.257 megajoules (mJ) for every kilogram of water vaporized. This use of energy for evaporation cools the soil, much the way it chills a person who comes out from the water after swimming on a windy day.

For example, if the amount of water associated with 100 g of dry soil was reduced by evaporation from 25 to 24 g (only about a 1% decrease) and if all the thermal energy needed to evaporate the water came from the moist soil, the soil would be cooled by about 12 °C. Such a figure is hypothetical because only a part of the heat of vaporization comes from the soil itself. Nevertheless, it indicates the tremendous cooling influence of evaporation.

The low temperature of a wet soil may be due partially to evaporation and partially to high specific heat. The temperature of the upper few centimeters of wet soil is commonly 3–6 °C lower than that of a drier soil. This is a significant factor in the spring in a temperate zone, when a few degrees will make the difference between the germination or lack of germination of seeds, or the microbial release or lack of release of nutrients from organic matter.

BOX 7.4

CALCULATING THE SPECIFIC HEAT (HEAT CAPACITY) OF MOIST SOILS

Soil water content markedly impacts soil temperature changes through its effect on the specific heat or heat capacity *c* of a soil. For example, consider two soils with comparable characteristics, soil A, a relatively dry soil with 10 g water/100 g soil solids, and soil B, a wetter soil with 30 g water/100 g soil solids.

We can assume the following values for specific heat:

$$\text{Water} = 1.0 \text{ cal/g and dry mineral soil} = 0.2 \text{ cal/g}$$

For soil A with 10 g water/100 g dry soil, or 0.1 g water/g dry soil, the number of calories required to raise the temperature of 0.1 g of water by 1 °C is: $0.1 \text{ g} \times 1.0 \text{ cal/g} = 0.1 \text{ cal}$. The corresponding figure for the 1.0 g of soil solids is:

$$1 \text{ g} \times 0.2 \text{ cal/g} = 0.2 \text{ cal}.$$

Thus, a total of 0.3 cal (0.1 + 0.2) is required to raise the temperature of 1.1 g (1.0 + 0.1) of the moist soil by 1 °C. Since the specific heat is the number of calories required to raise the temperature of 1 g of moist soil by 1 °C, we can calculate the specific heat of soil A as follows:

$$C_{\text{soil A}} = \frac{0.3}{1.1} = 0.273 \text{ cal/g}$$

These calculations can be expressed as a simple equation to calculate the weighted average specific heat of a mixture of substances:

$$c_{\text{moist soil}} = \frac{c_1 m_1 + c_2 m_2}{m_1 + m_2} \tag{7.7}$$

where c_1 and m_1 are the specific heat and mass of substance 1 (the dry mineral soil, in this case), and c_2 and m_2 are the specific heat of substance 2 (the water, in this case).

Applying this equation to soil A, we again calculate that $c_{\text{soil A}}$ is 0.273 cal/g, as follows:

$$c_{\text{soil A}} = \frac{0.2 \text{ cal/g} \times 1.0 \text{ g} + 1.0 \text{ cal/g} \times 0.10 \text{ g}}{1.0 \text{ g} + 0.10 \text{ g}}$$

$$= \frac{0.30 \text{ cal}}{1.1 \text{ g}} = 0.273 \text{ cal/g}$$

In the same manner, we calculate the specific heat of the wetter soil B:

$$c_{\text{soil B}} = \frac{0.2 \text{ cal/g} \times 1.0 \text{ g} + 1.0 \text{ cal/g} \times 0.30 \text{ g}}{1.0 \text{ g} + 0.30 \text{ g}}$$

$$= \frac{0.50 \text{ cal}}{1.3 \text{ g}} = 0.385 \text{ cal/g}$$

The wetter soil B has a specific heat c_B of 0.385 cal/g, whereas the drier soil A has a specific heat c_A of 0.273 cal/g. Because it must absorb an additional 0.112 cal (0.385−0.273) of solar radiation for every degree of temperature rise, the wetter soil will warm up much more slowly than the drier soil.

Thermal Conductivity of Soils

As shown in Section 7.9, some of the solar radiation that reaches the Earth slowly penetrates the profile largely by conduction; this is the same process by which heat moves to the handle of a cast-iron frying pan. The movement of heat in soil is analogous to the movement of water (see Section 5.5), the rate of flow being determined by a driving force and by the ease with which heat flows through the soil. This can be expressed as Fourier's law:

$$Q_h = \lambda \times \frac{\Delta T}{x} \tag{7.8}$$

where Q_h is the *thermal flux*, the quantity of heat transferred across a unit cross-sectional area in a unit time; λ is the **thermal conductivity** of the soil; and $\Delta T/x$ is the temperature gradient over distance x that serves as the driving force for the conduction of heat.

The thermal conductivity λ of soil is influenced by a number of factors, the most important being the moisture content of the soil and the degree of compaction (see Figure 7.36). Heat passes through water many times faster than through air. As the water content increases in a soil, the air content decreases, and the transfer resistance is decidedly lowered. When sufficient water is present to form a bridge between most of the soil particles, further additions will have little effect on heat conduction. Heat moves through mineral particles even faster than through water, so when particle-to-particle contact is increased by soil compaction, heat-transfer rates are also increased. Therefore, a wet, compacted soil would be the poorest insulator or the best conductor of heat. Here again the interconnectedness of soil properties is demonstrated.