GEOG 321 - Reading Package Lectures 10 & 11

The influence of the active surface extends down into a relatively shallow layer of the substrate (soil, snow, ice, bedrock). Because the active surface is the site of greatest energy absorption by day and depletion by night, it is also where the greatest thermal response is found. The effects diminish with distance away from the interface down into the substrate. The surface energy balance causes a temperature wave that establishes a temperature gradient $\partial T/\partial z$ and a sensible heat flux Q_G directed downwards by day and upwards at night. The size of both, $\partial T/\partial z$ and Q_G, diminish with distance away from the interface giving a changing vertical distribution (*profile*) of substrate temperatures illustrated in Figure 1.

In this section, we are confining ourselves to the semi-solid case of soil, and describe the interaction between the active surface and this soil layer due to heat and water movement. Soil temperatures vary less rapidly than air temperatures, and since radiative and convective exchanges are virtually absent in the soil, most of the activity takes place from molecule to molecule.

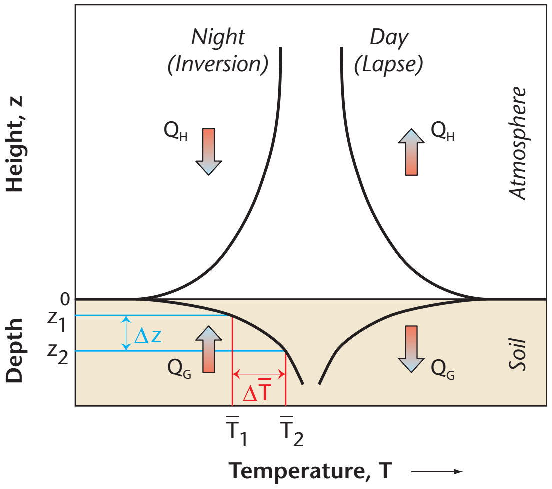


Figure 1: Idealized mean profiles of air and soil temperature and the sensible heat fluxes in atmosphere (Q_H) and in the soil (Q_G) in fine weather.

SOIL THERMAL PROPERTIES

The actual rate mean temperature change of a soil layer over time $\Delta T/\Delta t$ is proportional to the amount of heat currently absorbed by the layer (ΔQ) but it also depends on the thermal properties of the soil.

Heat capacity. When considering unit mass of the substance, the appropriate property is the specific heat (c, with units of $\mathrm{J\,kg^{-1}\,K^{-1}}$). Here, when dealing with unit volume, the property is the heat capacity (C, with units of $\mathrm{J\,m^{-3}\,K^{-1}}$), which is defined as the amount of heat (J) necessary to raise unit volume ($\mathrm{m^3}$) of a substance through a temperature change of 1 degree (K). Typical values of c and C are listed in Table 1 where it is clear that, compared to most other materials, water requires a large heat input to effect a given change in temperature whereas air requires very little. The heat capacity C relates to the ability of a substance to store heat and expresses the temperature change produced as a result of gaining or losing heat:

$$\frac{\Delta T}{\Delta t} = \frac{1}{C}\frac{\Delta Q}{\Delta z} \qquad \star \qquad (7.1)$$

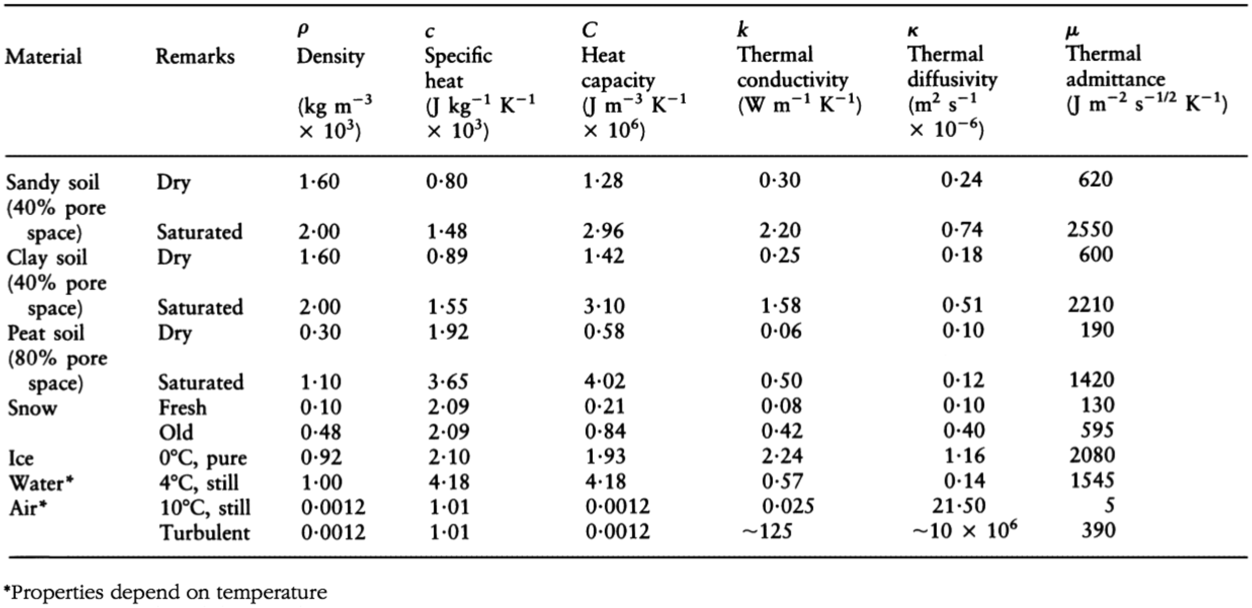
The value of C for a soil can be calculated by evaluating the fractions of soil solid, water and air. The value for a given soil is strongly dependent, in a linear fashion, on the soil moisture content (Figure 2a). Adding water with a very high heat capacity excludes a proportionate volume of soil air of low heat capacity (see Table 1). The result is a reduction in the soils thermal sensitivity.

The approximate value of the soil heat capacity can be obtained using simple soil analysis. The heat capacity of the soil is

$$C = \theta_m\,C_m + \theta_o\,C_o + \theta_w\,C_w + \theta_a\,C_a \qquad \star \qquad (7.2)$$

where θ_m, θ_o, θ_w and θ_a are the volume fractions of soil occupied by minerals, organic matter, water and air respectively.

Table 1: Thermal properties of natural materials



Material	Remarks	ρ Density ($\mathrm{kg\,m^{-3}} \times 10^3$)	c Specific heat ($\mathrm{J\,kg^{-1}\,K^{-1}} \times 10^3$)	C Heat capacity ($\mathrm{J\,m^{-3}\,K^{-1}} \times 10^6$)	k Thermal conductivity ($\mathrm{W\,m^{-1}\,K^{-1}}$)	κ Thermal diffusivity ($\mathrm{m^2\,s^{-1}} \times 10^{-6}$)	μ Thermal admittance ($\mathrm{J\,m^{-2}\,s^{-1/2}\,K^{-1}}$)
Sandy soil (40% pore space)	Dry	1·60	0·80	1·28	0·30	0·24	620
	Saturated	2·00	1·48	2·96	2·20	0·74	2550
Clay soil (40% pore space)	Dry	1·60	0·89	1·42	0·25	0·18	600
	Saturated	2·00	1·55	3·10	1·58	0·51	2210
Peat soil (80% pore space)	Dry	0·30	1·92	0·58	0·06	0·10	190
	Saturated	1·10	3·65	4·02	0·50	0·12	1420
Snow	Fresh	0·10	2·09	0·21	0·08	0·10	130
	Old	0·48	2·09	0·84	0·42	0·40	595
Ice	0°C, pure	0·92	2·10	1·93	2·24	1·16	2080
Water*	4°C, still	1·00	4·18	4·18	0·57	0·14	1545
Air*	10°C, still	0·0012	1·01	0·0012	0·025	21·50	5
	Turbulent	0·0012	1·01	0·0012	~125	~10 × 10^6	390

*Properties depend on temperature

Since C_a is very small compared with the other values the last term can be neglected without great error. θ_w can be found by oven-drying soil samples, θ_o can be obtained by firing-off such material in a furnace, leaving θ_m as the residual.

Example 1: Calculate the heat capacity of a dry mineral soil with a porosity of 55%. The heat capacity of the mineral soil matrix is $C_m = 2.1\,\mathrm{MJ\,m^{-3}\,K^{-1}}$.
Dry: $\theta_m = 0.45\,\mathrm{m^3\,m^{-3}}$, $\theta_o = 0$; $\theta_w = 0$

$$\begin{aligned} C &= C_m\theta_m = 2.1\,\mathrm{MJ\,m^{-3}\,K^{-1}} \times 0.45\,\mathrm{m^3\,m^{-3}} \\ &= \underline{0.9\,\mathrm{MJ\,m^{-3}\,K^{-1}}} \end{aligned}$$

Example 2: Calculate the heat capacity of the same soil when saturated.
Saturated: $\theta_m = 0.45\,\mathrm{m^3\,m^{-3}}$, $\theta_o = 0$; $\theta_w = 0.55$

$$\begin{aligned} C &= C_m\theta_m + C_w\theta_w \\ &= 2.1\,\mathrm{MJ\,m^{-3}\,K^{-1}} \times 0.45\,\mathrm{m^3\,m^{-3}} \\ &\quad +4.18\,\mathrm{MJ\,m^{-3}\,K^{-1}} \times 0.55\,\mathrm{m^3\,m^{-3}} \\ &= \underline{3.2\,\mathrm{MJ\,m^{-3}\,K^{-1}}} \end{aligned}$$

Thermal conductivity. Heat will flow from areas of high to lower temperature. The magnitude of the flux density (Q_G) will be proportional to the temperature difference over the layer. We can express this by Fouriers law:

$$Q_G = -k\frac{\partial T}{\partial z} \approx -k\frac{T_2 - T_1}{z_2 - z_1} \qquad \star \qquad (7.3)$$

where the subscripts refer to levels in the soil. This is a *flux-gradient form* of equation, i.e. the flux of an entity is equal to the ability to transfer times the gradient of a related property. Flux-gradient relations will re-occur throughout our discussions including other media than soil and other entities than heat. As before, the sign indicates the flux is in the direction of decreasing temperature. Therefore by day, when $\partial T/\partial z$ is negative the equation gives a positive value of Q_G; *k* is known as the *thermal conductivity* ($\mathrm{W\,m^{-1}\,K^{-1}}$), which is a measure of the ability to conduct heat. It is formally defined as the quantity of heat (J) flowing through unit cross-sectional area ($\mathrm{m^2}$) of the substance in unit time (s), if perpendicular to it there exists a temperature gradient of 1 $\mathrm{K\,m^{-1}}$. Typical values for a range of natural materials are listed in Table 1; of these motionless air is the most notable because it is such a very poor conductor of heat (i.e. it is a good insulator).

Unfortunately *k* is not a simple constant for a given soil. It varies both with depth and with time. However if we restrict ourselves to bulk averages *k* depends upon the conductivity of the soil particles, the soil porosity, and the soil moisture content. Of these the soil moisture content is the only short-term variable for a given soil. The addition of moisture to an initially dry soil increases its conductivity (Figure 2b). This happens for two reasons. First, coating the soil particles increases the thermal contact between grains. Second, since the soil pore space is finite the addition of pore water must expel a similar amount of pore air. From Table 1 we can see that this means replacing soil air with a substance whose conductivity is more than an order of magnitude greater.

Thermal diffusivity. The thermal diffusivity of a soil (κ, a greek 'kappa') is its ability to diffuse thermal influences. It controls the speed at which temperature waves move and the depth of thermal influence of the active surface:

$$\kappa = \frac{k}{C} \qquad \star \qquad (7.4)$$

which shows that thermal influences are directly proportional to the ability to conduct heat (k) but inversely proportional to the amount of heat necessary to effect temperature change (C). Thermal diffusivity may be viewed as a measure of the time required for temperature changes to travel. For example, the daytime heat input will generate a temperature wave that travels rapidly and to considerable depth in soil where conductivity is high, but if it takes large amounts of heat to warm intermediate layers because of a high heat capacity it will be slowed and not penetrate as far. Typical values of κ are given in Table 1.

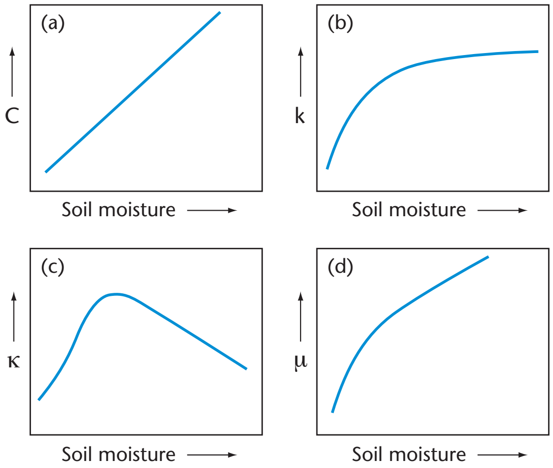


Figure 2: Relationship between soil moisture content: (a) thermal conductivity, (b) heat capacity, (c) thermal diffusivity and (d) thermal admittance for most soils.

The value of κ is obviously affected by the same soil properties that influence k and C, especially soil moisture (Figure 2c). Note that adding moisture to a dry soil initially produces a sharp increase in κ by increasing thermal contact and expelling soil air (i.e. increasing k as in Figure 2b). However, in most soils beyond about 20% soil moisture content by volume κ begins to decline. This happens because whereas k levels off (Figure 2b) the value of C continues to increase at higher moisture contents (Figure 2a).

Soils with high diffusivities allow rapid penetration of surface temperature changes and permit these effects to involve a thick layer. Thus for the same heat input their temperature regimes are less extreme than for soils with low diffusivities. By day the surface heating is used to warm a thick layer of soil, and at night the surface cooling can be partially offset by drawing upon heat from a similarly thick stratum. Soils with poor diffusivities concentrate their thermal exchanges only in the uppermost layer, and consequently experience relatively extreme diurnal temperature fluctuations. Therefore, in general a wet clay has a conservative thermal climate, whereas an almost dry peat is extreme.

Thermal admittance. This is a convenient juncture to introduce a related thermal property called the soil thermal admittance (μ):

$$\mu = \sqrt{kC} \qquad \star \qquad (7.5)$$

Technically this is a surface rather than a soil property. It is a measure of the ability of a surface to accept or release heat since it expresses the temperature change produced by a given heat flux change. This is why some materials with high μ (such as metals) initially feel cooler to the touch than those with low μ (such as wood) even though both are actually at the same room temperature of, say, 20°C. Your finger is at a temperature of about 30°C and the 10-degree temperature difference is rapidly sensed at a surface of high μ because heat transfer at the finger- material interface is excellent. Similarly, note the apparent difference of placing your bare foot on a clay tile bathroom floor compared with a bathmat or cork tile. On the other hand, if the materials being touched are at a higher temperature than your hand or foot, that with higher μ would initially feel warmer.

A surface also has an analogous atmospheric admittance ($\mu_a = \sqrt{C_a k_a}$). The relative magnitude of the two properties are important in determining the sharing of sensible heat between the soil and the atmosphere since:

$$\frac{\mu_s}{\mu_a} = \frac{Q_G}{Q_H} \tag{7.6}$$

where the subscript s refers to the soil and the subscript a to the atmosphere. The amplitude of the surface temperature wave is closely linked to these properties. The larger these values are the easier it is for heat to be transported to or from the interface and the smaller will be the surface temperature variations. For a given site the value of μ_a is determined by the state of turbulence (i.e. the eddy diffusivity, K_H), and μ_s by the soil moisture (Figure 2d). For a given state of the atmosphere, sites with large μ_s will accept or release heat to or from soil storage with relative ease and hence will exhibit relatively small surface temperature changes through a day.

SOIL TEMPERATURES

The course of soil temperature at a particular depth T_z with time is very regular in comparison with almost any other atmospheric element. Typical soil temperature variations at a number of depths on a cloudless day are given in Figure 3. The near- surface temperature variation is wave-like and agrees closely with that of the surface. The wave penetrates downward to lower depths, but in doing so its amplitude decreases, and the times of maximum and minimum temperature are lagged (phase shift to the right in time). Both features depend on κ_s. The wave amplitude at any depth is given by:

$$\Delta\overline{T}_z = \Delta\overline{T}_0 \exp\left(-z\sqrt{\frac{\pi}{\kappa P}}\right) \tag{7.7}$$

where, $\Delta\overline{T}_0$ – surface temperature wave amplitude, P – wave period (s). This shows that the diurnal temperature range decreases exponentially with depth. In most soils the daily surface temperature wave is only discernible to a depth of about 0.75 m. In soils with low κ values it is even less, indicating that flux convergence has extinguished Q_G in a thin near-surface layer. The time lag (phase shift) for the wave crest (maximum) and trough (minimum) to reach lower depths is given by:

$$(t_2 - t_1) = \frac{z_2 - z_1}{2}\sqrt{\frac{P}{\pi\kappa}} \tag{7.8}$$

where t_1 and t_2 are the times at which the wave crest or trough reaches depths z_1 and z_2. Note that because of this time lag, at any given time the soil may be cooling in its upper layers but warming at only a short distance beneath, and vice versa when the upper layers are warming (Figure 3a).

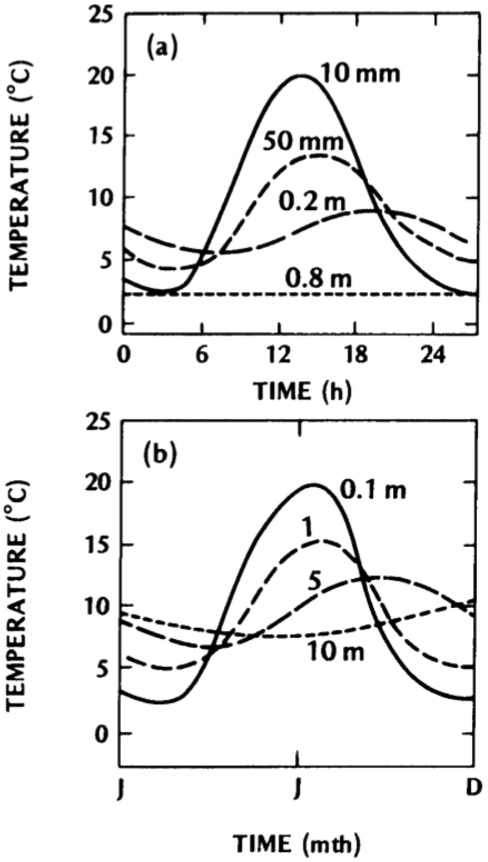


Figure 3: Generalized cycles of soil temperature at different depths for (a) daily and (b) annual periods.

The annual soil temperature regime (Figure 3b) follows a wavelike pattern entirely analogous with the diurnal one. The wave period is of course dependent upon the annual rather than the daily solar cycle. With that adjustment, equations 2.6 and 2.7 still apply to the wave amplitude and lag with depth. With a longer period the wave amplitude decreases less rapidly with depth, and the depth of the affected layer is much greater than for the diurnal case. In fact the difference between the depth of penetration of the annual and daily waves is approximated by the square root of the ratio of their respective periods, i.e. $\sqrt{3.15 \times 10^7 \mathrm{s} / 8.64 \times 10^4\,\mathrm{s}} = 19$. Therefore given the daily depth is 0.75 m in typical soils, the annual wave may penetrate to about 14 m. This would then be termed the depth of zero annual range. The temperature at this depth is sometimes used as a surrogate for the average annual air temperature of the site. This is based on the premise that long-term thermal equilibrium exists between the soil and the atmosphere. During the warm season soil temperatures decrease with depth and the associated downward heat flux builds up the soils heat store. In the cold season the gradient is reversed and the store is gradually depleted. The spring and autumn are transitional periods when the soil temperature gradient reverses sign. These reversals (or 'turnovers') are important biological triggers to soil animals and insects. In the spring they may come out of hibernation, and/or move upwards towards the warmer surface layers. In the autumn they retreat to depths where soil warmth is more equable.

The effects of cloud on the diurnal soil temperature pattern are fairly obvious. With overcast skies absolute temperatures are lower by day but warmer at night, and the wave amplitude is smaller; variable cloudiness induces an irregular pattern upon the diurnal wave. Using the same argument as for the annual/daily comparison a 15 min temperature variation induced by cloud would only travel about one tenth of the distance of the daily wave, so typically it would not be registered below about 75 mm. Rainfall is capable of either increasing or decreasing soil temperatures depending upon the temperature of the rain in comparison with the soil. It is also capable of transporting heat as it percolates down through the soil.

SOIL WATER FLOW AND SOIL MOISTURE

Soil moisture is usually expressed in one of two ways. Soil moisture content (θ_w) is a measure of the actual water content, and is defined as the percentage volume of a moist soil occupied by water. This is particularly pertinent in water balance studies where changes in mass are important. Soil moisture potential (Ψ) on the other hand is an indirect measure of water content, and may be visualized as the energy necessary to extract water from the soil matrix. The units of Ψ are those of negative pressure (Pa), which can also be expressed as a head of water displaced (1 m head of water = 1.0×10^4 Pa). This concept is of value in estimating the availability of water for plant use, and in calculating moisture movement.

The forces which bind soil water are related to the soil porosity and the soil water content. The forces are weakest in the case of open textured, wet soils, and greatest for compact dry soils. Thus at a given value of θ_w the water potential is greatest for a clay, least for a coarse sand, and intermediate for a loam.

Measuring soil moisture. Soil moisture and its change over time can be measured by simple gravimetric methods or time-domain reflectometry (TDR, see lecture slides). The *gravimetric method* involves direct sampling using an auger to remove soil from the required depths. The weight loss after oven-drying at 105°C gives the water content by mass (Ω_w in $\mathrm{kg\,kg^{-1}}$), and knowledge of the soil's bulk density (ρ_s, in $\mathrm{kg\,m^{-3}}$) allows calculation of the soil water content by volume (θ_w, in $\mathrm{m^3\,m^{-3}}$):

$$\theta_w = \frac{\rho_s\,\Omega_w}{\rho_w} \qquad (7.9)$$

where ρ_w is the density of water.