Lecture 34, Soil Physics, Theory and Observations of Heat Transfer Part 1

November 21, 2014

Instructor: Dennis Baldocchi
Professor of Biometeorology
Ecosystem Science Division
Department of Environmental Science, Policy and Management
345 Hilgard Hall
University of California, Berkeley
Berkeley, CA 94720

Lecture Topics

1. Soil Properties
   a. soil texture and classes
   b. heat capacity, conductivity and thermal diffusivity
   c. moisture conductivity
   d. porosity and diffusion

2. Theory, Heat Transfer
   a. Heat transfer, Fourier’s equation

3. Observations
   A. Temperature profiles
      i. Daily patterns, Phase lags and Amplitudes of soil temperature
      ii. Seasonal Patterns
   B. Soil Heat Flux
      a. Daily patterns
      b. Seasonal Patterns

34.1 Introduction

“Darkle, darkle, little grain,
I wonder how you entertain
A thousand creatures microscopic.
Grains like you from pole to tropic
Support land life upon this planet
I marvel at you, crumb of granite! “

Dr. Francis Hole, Soil Scientist (1913-2002)

The soil is an important component of biometeorology. The soil surface is the lowest boundary of the biosphere-atmosphere system. Rules and tools learn so far for the atmosphere, need to be modified when applied to the soil. The soil environment contrasts
vastly with the atmosphere. It is a porous medium. Air flow in the soil is laminar or diffusive. And it has a greater heat capacity and density than air, so it alters how heat is transferred between the biosphere and atmosphere.

It is a complex medium, being a reservoir of roots, water, organic matter, microbes, soil fauna, nutrients, minerals and gases. The relative proportions of the soil, liquid and gas constituents differ vastly from what is observed in the free atmosphere. In an unsaturated state, the soil contains matter in three states of matter, solid, gas and liquid. The solid aspects of the soil include minerals, organic matter (roots, litter, peat, microbes, fauna). Water is the major liquid component of soils. The air segment contains oxygen, nitrogen, CO₂, water vapor, nitrous oxide, and methane. In a dry state it is a porous medium, containing air pockets of various size, depending on the soil texture, and macropores, formed by decayed roots, worms and insects. The unsaturated soil layer is called the vadose zone.

The amount of air will change with the amount of moisture in the soil pores. The composition of air in the soil will also be very distinct from that of the atmosphere, as roots and aerobic microbes consume oxygen and exhale carbon dioxide and nitrous oxide.

1. Properties of Soil

As one digs into a column of soils one will find a distinct and stratified profiles. The soil profile consists of five distinct layers, called horizons (O,A,B,C and R). The top-most horizon is the organic layer. It has two sublayers. Decomposing vegetative debris is in O1 and decomposing plant and animal debris is in O2. Horizons A and B are the ‘true’ soil horizons. The A horizon contains soil mineral. Leaching from excess rain may have depleted the horizon of clay, iron and aluminum oxides. The underlying B horizon retains minerals washed out of the A horizon. The soil parent material is associate with the C horizon. Little biology occurs in this zone. Finally, there is the R horizon, the bedrock layer.

Textural classes of soil include gravel, sand, silt and clay. Particle size diminishes with each texture. Gravel is larger than 2 mm, sand ranges between 2 and 0.05 mm, silt between 0.05 and 0.005 mm and clay has the smallest particles.

Another descriptive division is soil class, which are comprised of a mixture of clay, sand and silt. A well-mixed soil is a loam, with about 40% sand and silt and 20% clay.
There are a variety of measures used to define and characterize soils. Particle density is the mass of a soil particle per unit volume, as opposed to bulk density, the mass of a soil sample per unit volume. Particle density of quartz type minerals is 2650 kg m$^{-3}$.

Clays are composed of crystals with layers of silicon, aluminum and oxygen atoms. Important clay formations are kaolinite, illite, vermiculite and montmorillonite.

**Bulk density** the ratio of the mass of the solid in a known volume. The mass of the soil is equal to the product of the solid density times its volume:

$$\rho_b = \frac{m_s}{v_t} = \frac{v_s \rho_s}{v_t}$$

Total Porosity is the ratio of volume of air, $v_a$, and water, $v_w$, to sum of air, water and solid, $v_s$

$$P = \frac{v_a + v_w}{v_a + v_w + v_s}$$
It can also be computed as

\[ P = 1 - \frac{\rho_b}{\rho_s} \]

The gas filled porosity is the difference between the total porosity and the volumetric water content

\[ P_g = \phi = \frac{V_{\text{air}} + V_{\text{water}}}{V_{\text{solid}} + V_{\text{air}} + V_{\text{water}}} = 1 - \frac{\rho_b}{\rho_s} - \theta \]

**Measures of Soil Moisture**

Mass ratio, Mass of water to mass of solid

\[ w = \frac{m_w}{m_s} = \frac{v_w \rho_w}{v_s \rho_s} = \theta \frac{\rho_b}{\rho_i} \]

This method is measured via the classic gravimetric sampling. Take a soil core, weigh it wet, weigh it dry (at 105 °C) and compute the ratios of evaporated water to dry soil remaining.

Though easy to measure, gravimetric water content is difficult to implement when interpreting soil and plant processes, which scale better volumetric water content.

**Volumetric water content**

\[ \theta = \frac{v_w}{v_i} = w \frac{\rho_b}{\rho_w} \]

As you can see, volumetric and gravimetric soil moisture contents differ by the bulk density of the soil, \( \rho_b \)

After Tyndall et al.. 1999

<table>
<thead>
<tr>
<th>Soil type</th>
<th>Bulk density, kg m-3</th>
<th>Porosity, percent</th>
</tr>
</thead>
<tbody>
<tr>
<td>Peat</td>
<td>0.65-1.1</td>
<td>60-80</td>
</tr>
<tr>
<td>Ideal soil</td>
<td>1310</td>
<td>50-60</td>
</tr>
<tr>
<td>Clay</td>
<td>1220</td>
<td>45-55</td>
</tr>
<tr>
<td>Silt</td>
<td>1280</td>
<td>40-50</td>
</tr>
<tr>
<td>Medium to coarse sand</td>
<td>1530</td>
<td>35-40</td>
</tr>
</tbody>
</table>
Uniform sand | 1650 | 30-40
Fine to medium sand | 1850 | 30-35
Gravel | 1870 | 30-40

Hence, the volumetric water content can be computed by measuring the gravimetric soil moisture content and the bulk density of the soil.

**Heat transfer equation**

Heat transfer into and out of soil can be computed using the Fourier Heat transfer equation. For the one dimensional case, the time rate of change of soil temperature, $T$, is a function of the flux divergence of soil heat flux into or out of the soil, $G$:

$$\rho_s C_s \frac{\partial T}{\partial t} = - \frac{\partial G}{\partial z}$$

If we express $G$ in terms of a Fickian equation, $G = k \frac{\partial T}{\partial z}$, then we can reexpress the heat transfer equation as a function of soil temperature, $T$:

$$\rho_s C_s \frac{\partial T}{\partial t} = - \frac{\partial}{\partial z} (k \frac{\partial T}{\partial z})$$

where $k$ is thermal conductivity, $\rho_s$ is the density of the soil and $C_s$ is the specific heat of the soil. Normalization produces:

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2}$$

Thermal diffusivity is $\kappa = \frac{k}{\rho_s C_s}$

Energy exchange associated with latent heat of evaporation or phase changes due to condensation are ignored here.

**Heat Capacity**

The heat capacity of soil is a weighted quantity, representing the relative proportions of water, mineral and organic matter. It is a function of the respective specific heat capacities ($C$), the density of the constituent ($\rho$) and the volume fraction in the soil ($\theta, \phi$):

$$\rho_s C_s = \phi_m \rho_m C_m + \phi_o \rho_o C_o + \theta_w \rho_w C_w$$
Campbell (Campbell, 1985) computes $C_s$ as a weighted fraction of the bulk density and water content:

$$C_s = \frac{2400000 \rho_b}{2.65} + 4180000 \theta$$

When the soil is frozen the above equation is modified to consider this component.

![Soil Moisture content](image)

Figure 2 Adapted from Campbell and Norman (1998)

<table>
<thead>
<tr>
<th>material</th>
<th>Density (Mg m$^{-3}$)</th>
<th>Specific Heat (J g$^{-1}$ K$^{-1}$)</th>
<th>Thermal conductivity (W m$^{-1}$ K$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil minerals</td>
<td>2.65</td>
<td>0.87</td>
<td>2.5</td>
</tr>
<tr>
<td>Granite</td>
<td>2.64</td>
<td>0.82</td>
<td>3.0</td>
</tr>
<tr>
<td>Quartz</td>
<td>2.66</td>
<td>0.80</td>
<td>8.8</td>
</tr>
</tbody>
</table>
Thermal Conductivity

Thermal conductivity depends upon the relative fractions of mineral, water, gas and organic matter.

Campbell and Norman (Campbell and Norman, 1998) present a weighted relation, with origins attributed to de Vries

\[
k_{\text{soil}} = \frac{\phi_{w} \xi_{w} k_{w} + \phi_{g} \xi_{g} k_{g} + \phi_{m} \xi_{m} k_{m}}{\phi_{w} \xi_{w} + \phi_{g} \xi_{g} + \phi_{m} \xi_{m}}
\]

units W m\(^{-1}\) K\(^{-1}\). The weighting factors, \(\xi\), are introduced to consider heat associated with evaporation and condensation.

The latent heat transfer through soil pores is estimated as:

\[
\lambda E = -\lambda \rho D_{v} \frac{dc_{v}}{dz} = -\lambda \rho D_{v} \frac{dc_{v}}{dT} \frac{dT}{dz}
\]

\(\rho\) is the density of air, \(D_{v}\) is the molecular diffusivity of vapor, and \(c_{v}\) is the mole fraction of vapor. A simpler version of the evaporation equation can be assessed as:

\[
\lambda E = -\lambda \rho D_{v} s h_{r} \frac{dT}{dz}
\]

where \(s\) is the slope of the saturation vapor pressure-temperature curve and \(h_{r}\) is relative humidity.
Figure 3 Computations of thermal conductivity as a function of soil texture and volumetric water content

The thermal conductivity of gas is the sum of the dry air component and the vapor in the soil pores.

\[ k_g = k_a + \frac{\lambda s h_r f_w \rho D_v}{p_a - e_a} \]

- \( D_v \) is the molecular diffusivity of water vapor
- \( \lambda \) is the latent heat of evaporation
- \( p_a \) is a pressure
- \( e_a \) is vapor pressure
- \( h_r \) is the relative humidity of the soil
- \( \rho \) is the molar density of air
- \( s \) is the slope of the saturation vapor pressure-temperature curve
- \( f_w \) is a flow factor

The fluid conductivity is:

\[ k_f = k_g + f_w (k_w - k_g) \]
An engineering relation is used by Campbell (Campbell, 1985) to compute thermal diffusivity for a variety of soil types

\[ K = A + B \theta - (A - D) \exp(-C \theta^E) \]

A, B and D are functions of bulk density:

\[ A = 0.65 - 0.78 \rho_b + 0.60 \rho_b^2 \]

\[ D = 0.03 + 0.10 \rho_b^2 \]

B is also dependent on the volumetric water content

\[ B = 1.06 \rho_b \theta \]

C is a function of the clay fraction (c):

\[ C = 1 + 2.6c^{-1/2} \]

E is 4.

For dry soils soil radiation is portioned into heating the soil and air. But what and how is this energy partitioned??

\[ R_n = G + H \]

The ratio between G and H is a function of their admittances. For soil the admittance is the square root of the product of its thermal conductivity and volumetric heat capacity

\[ \frac{G}{H} = \frac{\mu_{\text{soil}}}{\mu_{\text{air}}} = \sqrt{\frac{k_{\text{soil}} \rho_{\text{soil}} C_{\text{soil}}}{k_{\text{air}} \rho_{\text{air}} C_{\text{air}}}} \]

K air will vary for turbulent, laminar and still air

When admittance is high available energy goes into heating the soil. When it is low, the air is heated.

Values of G/H (Campbell and Norman)
<table>
<thead>
<tr>
<th>Medium</th>
<th>Bare, dry soil</th>
<th>Loose straw much</th>
</tr>
</thead>
<tbody>
<tr>
<td>Still air</td>
<td>50</td>
<td>20</td>
</tr>
<tr>
<td>Calm atmosphere</td>
<td>0.5</td>
<td>0.2</td>
</tr>
<tr>
<td>Windy atmosphere</td>
<td>0.1</td>
<td>0.04</td>
</tr>
</tbody>
</table>

**Heat transfer properties**

Phase lags, Amplitudes, Damping Depth and thermal diffusivity can be deduced from soil temperature measurements

\[ D = \sqrt{\frac{2\kappa}{\omega}} \]

\[ \bar{\sigma} = \frac{2\pi}{\tau} \]

\( \omega \) is angular frequency (radians per second)

\( \tau \) is period of fluctuation

\[ T(z,t) = \bar{T} + A(0) \exp(-z/D) \sin(\bar{\sigma}(t - t_0) - z/D) \]

**Thermal Diffusivity**

Several methods can be used to deduce thermal from field measurements. Verhoef et al. (Verhoef et al., 1996) have tested the Amplitude equation, the Phase Equation, the arctangent method and the logarithmic equation. For simplicity we introduce the first two methods

**Amplitude Equation**

\[ \kappa = \frac{\omega}{2} \left[ \frac{z_2 - z_1}{\ln(A_1/A_2)} \right]^2 \]

where A is the amplitude of the temperature course at depth z.

**Phase Equation**
\[ \kappa = \frac{1}{2 \omega} \left[ \frac{z_2 - z_1}{\Delta t} \right]^2 \]

\( \Delta t \) is the time interval of the maximum soil temperature at depths \( z_1 \) and \( z_2 \).

3. Soil Temperature Profile

Figure 4 Measurements of temperature profiles under ponderosa pine over the course of a summer day

Diurnal patterns of soil temperature
The diurnal pattern of soil temperature varies with depth, in terms of amplitude and phase. The dark soil under a closed forest canopy varies only 3 degrees. The amount of energy reaching the soil surface has a marked impact on its soil temperature regime.

Figure 5 Diurnal pattern of soil temperature under a temperate broadleaved forest on a summer day. Note changes in phase and amplitude with depth.

This small amplitude of soil temperature under a deciduous forest is perturbed when the forest is leafless
The implications on choosing the correct soil temperature to drive soil mediated processes like soil respiration are huge. We hence advocate the measurement of soil profiles, logarithmically spaced that corresponds with depths of biological activity, e.g., roots, location of microbes etc.

In contrast, an open ponderosa pine stand experiences a swing of 25°C over the course of a day near the surface. In our study during the summer of 2000, soil temperatures at the surface often approached 0°C at night and 50°C during the day!

Figure 6 Diurnal course of soil temperature under a leafless deciduous forest
Figure 7 Soil temperature course under a sparse pine plantation. Note how surface temperature almost reaches freezing early in the morning and is over 45 °C by midday, a 30 °C ++ diurnal range! Also compared to the grassland, the diurnal curve is not smooth, as parts of the ground are shaded, or not by trees and exposed in gaps.

Collecting data from a California grassland shows.
Figure 8 Soil Temperature data Vaira Grassland, Day 204. Dry soil, dead grass, clear skies. The overlying vegetation is more uniform, so is the diurnal course of soil temperature.

Using the phase equation we deduce that the thermal diffusivity is $0.377 \text{ mm}^2 \text{ s}^{-1}$, a value representative of a dry soil, as when soil moisture was less than 10%. In contrast, $k$ is $1.13 \text{ mm}^2 \text{ s}^{-1}$ on day 56, during the wet winter when soil moisture was about 30%.

Additional information on relations between thermal diffusivity and soil moisture is shown by Verhoef et al. (Verhoef et al., 1996).
The annual course of soil temperature

The annual course of air and soil temperature is examined next.
Figure 10 seasonal trend in air and soil temperature deciduous forest

Note how average deep soil temperature converges with annual air temperature at a range of climates. On this topic, data was culled from the Fluxnet data base to examine the relation between annual soil and air temperature. The correlation was quite high.

Figure 11 Soil temperature is an excellent proxy for mean climate data and climate change
Why is this relation important? It can help us prescribe lower boundary conditions for SVAT models. We need to know the deep soil temperature, we can assign it based on climate data.

Soil temperature is a key function of soil moisture. Soils will become much warmer as they dry. Cognition of this fact is critical for modeling soil respiration, independent of soil moisture. One will overestimate Q10 relations of respiration as the soil dries and the soil temperature range grows.

![Graph showing soil temperature at 10 cm depth vs. soil volumetric moisture](image)

**Fig. 5** Soil temperature (10 cm) and soil moisture are negatively correlated at high soil moisture conditions. Data points are averages across all measurement locations.

**Figure 12 Xu and Qi**

**Soil Heat Flux Density**

In order to achieve information on the biologically relevant temperature and moisture content for soil respiration it is imperative that we are able to model mass and energy exchange at the soil. The following figures show data on soil heat flux density.
Daily course of Soil Heat Flux Density

1. A crop, LAI of about 3, relatively open system

- Figure 13: Diurnal pattern of $G$ under a closed wheat canopy.

2. Diurnal pattern of soil heat flux under a closed canopy deciduous forest. Note that the range of values is much smaller.

forest
Figure 14 Diurnal pattern of $G$ under a closed deciduous forest canopy

Annual course of soil heat flux density
Figure 15 Daily averaged seasonal course of G under a deciduous forest.

Note how the system loses G in the winter and gains it in the summer. The annual amplitude is small (20 W m$^{-2}$) and averages to about zero over the course of a year.

Discussion Topic

Can a network of soil heat flux plates and deep soil temperature provide an objective measure of global warming?

There is a Soil Climate Analysis Network operating in the US

http://www.wcc.nrcs.usda.gov(scan/)
Summary Points

Soils consist of a mixture of silt, sand and clay and organic matter. Together, these components affect the soil’s texture, water holding capacity and heat capacity and conductivity.

Sand has the highest heat capacity and conductivity and litter/organic matter the lowest, when dry. Heat capacity and conductivity of all soils increase in a non-linear manner with soil moisture.

The amplitude of the temporal course of temperature (daily/annual) decreases and becomes more lagged (relative to air temperature) with soil depth.

References:


EndNote