Soil Moisture

Even though the topic may not seem like a usual topic of meteorology, but rather a part of the extensive field of *hydrology*, it is important for atmospheric scientists to have some basic idea about the soil processes. Soil moisture is important for the atmosphere for many reasons. Soil is where plants get their water, and, hence, it directly affects the transpiration by the vegetation. Over bare soils, evaporation rate directly depends on soil moisture. The moisture can affect the surface albedo; for example, moist soils are usually darker in color than dry soils. Moist soils usually conduct and store heat much better than dry soils. Etc, etc. Thus, to develop comprehensive cloud-resolving, regional or large-scale atmospheric models able to reproduce diurnal cycle and boundary layer development, it is important to represent and understand soil processes. These processes are quite complex, and only minimum introduction will be given here.

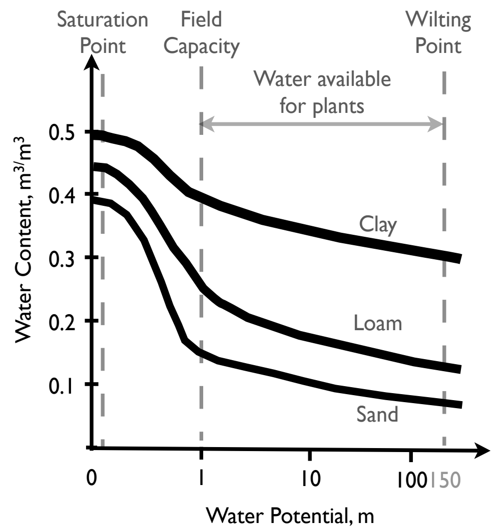
One of the most fundamental concepts related to the ground water is the *water potential* ψ. Conceptually, it is easy to understand by a simple notion that free water usually runs downhill, that is from high to low gravitational potential. Underground water also 'wants' to go from high to low potential. However, underground, in addition to gravitational potential $\rho_w gz$, there is an attraction of water to soil particles, which greatly restricts the movement of water in soils. Gravitational potential has unites of energy per unit area. However, because of the constant factor $\rho_w g$, water potential is usually measured in units of height (meters) or pressure (bars) using a simple unit converging rule that 1 bar of pressure is created by 10-m column of water.

The water potential is generally viewed as the sum of three terms: $\psi = \psi_g + \psi_m + \psi_o$. The gravitational potential ψ_g, mentioned above, is simply the height above some reference level. It can be negative or positive. The exact reference level is not that important; what really matters is the vertical gradient of ψ_g, which simply equals to unity if potential is measured in meters. The potential ψ_m is called *matric or matrix potential*. It describes soil's attraction or adhesion to water. The adhesion depends on soil's fine structure and composition called soil matrix. Generally, the smaller soil particles, the stronger the attraction and it is more difficult to move water through. Because it always takes energy to move water through soil, ψ_m is always negative. So adding matric potential is equivalent to lowering the gravitational potential of the soil water. The dryer the soil, the higher adhesion of water to the soil particles, the more negative the matric potential becomes and the higher suction force should be applied by a plant root system to extract water from the soil. Finally, the potential ψ_o is called the *osmotic potential*. It represents the effects of dissolved salts in the soil. Out of three potentials, the matric potential ψ_m is the most important for the soil water movement in unsaturated conditions, that is, when there is no standing water or muddy soil.

The quantitative measure of the soil moisture is a *volumetric moisture content*, η, or simply moisture content. It is defined as a ratio of the volume of soil water to the volume of soil. Different types of soil have different key characteristics in terms of water content as shown by the table below.

For each soil type, there is a unique relationship between moisture content and matrix potential as shown by the figure. Note that the positive values represent absolute values as matrix potential is generally negative.

Saturation point is achieved when water is freely running through soil, so that all soil pore spaces are filled with moving water, so that all soil air is pushed out. Generally, this water is considered to be unavailable to plants to take up by their root system. Also, the soil in that state lacks oxygen needed for the roots. This condition is typical for flood. Saturation points of various soil types are rather similar. The water potential for that soil condition is generally less than -1 m as shown by the table below.



Field capacity is achieved when flood is over and water stops running freely through the soil matrix with the soil pores are partially filled with air. It is the maximum water content that soil can hold without water running down by the force of gravity. Note that for the soils with coarser particles (sand, loam), the field capacity content can be substantially lower than saturated value, while for soils composed of finer particles (clays), the field capacity is similar to saturation point. The soil potential corresponding to field capacity is about -1 m.

Wilting point is the water content so low that the adhesion of remaining water to the soil particles is stronger than plants' ability to extract it and transpire. Below the wilting point the plants wilt, which may lead to plants' death. Note that there could be still plenty of water in the soil at wilting point (see the figure and table), but it is generally unavailable to plants. Note that water past the wilting point can still make it's way to the soil surface and contribute to evaporation.

Table A9. *Soil moisture quantities for a range of soil types, based on Clapp and Hornberger* (1978)

Quantities shown are as follows: η_s is the saturation moisture content (volume per volume), η_w is the wilting value of the moisture constant which assumes 150 m suction (i.e. the value of η when $\psi = -150$ m), ψ_s is the saturation moisture potential and $K_{\eta s}$ is the saturation hydraulic conductivity; b is an index parameter (see Eqs. 5.46–5.48).

Soil type	η_s ($m^3\ m^{-3}$)	η_{fc}	ψ_s (m)	$K_{\eta s}$ (10^{-6} m s^{-1})	b	η_w ($m^3\ m^{-3}$)
1. sand	0.395	0.15	− 0.121	176	4.05	0.0677
2. loamy sand	0.410	0.18	− 0.090	156.3	4.38	0.075
3. sandy loam	0.435	0.20	− 0.218	34.1	4.90	0.1142
4. silt loam	0.485	0.30	− 0.786	7.2	5.30	0.1794
5. loam	0.451	0.25	− 0.478	7.0	5.39	0.1547
6. sandy clay loam	0.420		− 0.299	6.3	7.12	0.1749
7. silty clay loam	0.477	0.38	− 0.356	1.7	7.75	0.2181
8. clay loam	0.476	0.40	− 0.630	2.5	8.52	0.2498
9. sandy clay	0.426		− 0.153	2.2	10.40	0.2193
10. silty clay	0.492	0.40	− 0.490	1.0	10.40	0.2832
11. clay	0.482	0.40	− 0.405	1.3	11.40	0.2864

Here is a simple formula that relates the water potential to the soil water content:

$$\psi = \psi_s \left(\frac{\eta}{\eta_s} \right)^{-b} \qquad (1)$$

where ψ_s, η_s, and b are soil parameters that depend only on soil type (see the table). Evaporation from the bare soil surface is given by the modified bulk aerodynamic formula:

$$E = \rho (R_h q_*(T_s) - q) / r_a \qquad (2)$$

where R_h is the relative humidity at the soil surface. It depends on the surface soil water potential ψ_{sfc} as

$$R_h = \exp(g \psi_{sfc} / R_v T_s) \qquad (3)$$

The ψ_{sfc} can be computed from the surface soil water content using (1). The dependence of R_h and evaporation rate on soil moisture content is shown in the figure below (adopted from Garratt).

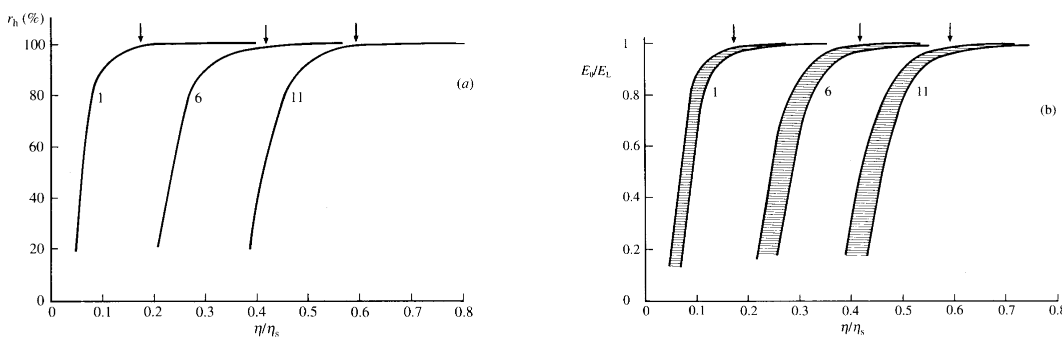


Fig. 5.9 (*a*) Relative humidity r_h as a function of relative soil moisture content η/η_s, based on Eq. 5.49 and data in Table A9 for soil types 1 (sand), 6 (loam) and 11 (clay). Calculations are for a temperature T_0 of 303 K. The vertical arrows indicate the wilting points. Note that combining Eqs. 5.46 and 5.49 allows r_h to be calculated from $\ln r_h = -(g/R_v T_0)\psi_s(\eta/\eta_s)^{-b}$. (*b*) E_0/E_L as a function of the relative soil moisture content, based on numerical simulations in an atmospheric model for a range of climate conditions (mid-latitude summer) represented by the shaded regions (the temperature range is 283–303 K and $q = 0.005$).

The movement of water through the soil is parameterized using a variant of the Darcy's law that states that mass flux of water F_w through the soil is proportional to the vertical gradient of total water potential (the z axis is assumed pointing up):

$$F_w = -\rho_w K(\eta)\frac{\partial}{\partial z}(z+\psi) \qquad (4)$$

where $K(\eta)$ is hydraulic conductivity coefficient in units of velocity (m/s). It depends on soil water content as

$$K(\eta) = K_s\left(\frac{\eta}{\eta_s}\right)^{2b+3} \qquad (5)$$

where saturation value K_s is the coefficient for saturated conditions given by the table above. Note that because the values of *b* for various soil types are generally in a wide range from 4 to 11 (see the table), the sensitivity of ψ, R_h and $K(\eta)$ to changes in soil content can be very high. Finally, conservation of water dictates that

$$\rho_w\frac{\partial\eta}{\partial t} = -\frac{\partial F_w}{\partial z} \qquad (6)$$

Note that for the surface soil layer, one would also add the precipitation source and also the surface run-off sink of water. At the bottom layer, one would need to consider the drainage of water to the water table. Depending on the depth of the root system, one would need to add a sink due to uprooting of water by plants.