Lecture 11. Surface evaporation and soil moisture (Garratt 5.3)

In this lecture…

- Partitioning between sensible and latent heat fluxes over moist and vegetated surfaces
- Vertical movement of soil moisture
- Land surface models

Evaporation from moist surfaces

The partitioning of the surface turbulent energy flux into sensible vs. latent heat flux is very important to the boundary layer development. Over ocean, SST varies relatively slowly and bulk formulas are useful, but over land, the surface temperature and humidity depend on interactions of the BL and the surface. How, then, can the partitioning be predicted?

For saturated ideal surfaces (such as saturated soil or wet vegetation), this is relatively straight- forward. Suppose that the surface temperature is T_0. Then the surface mixing ratio is its saturation value $q^*(T_0)$. Let z_1 denote a measurement height within the surface layer (e. g. 2 m or 10 m), at which the temperature and humidity are T_1 and q_1. The stability is characterized by an Obhukov length L. The roughness length and thermal roughness lengths are z_0 and z_T. Then Monin-Obuhkov theory implies that the sensible and latent heat fluxes are

$$H_S = \varrho L_v C_H V_1 (T_0 - T_1),$$

$$H_L = \varrho L_v C_H V_1 (q_0 - q_1), \text{ where } C_H = \mathrm{fn}(V_1, z_1, z_0, z_T, L)$$

We can eliminate T_0 using a linearized version of the Clausius-Clapeyron equations:

$$q_0 - q^*(T_1) = (dq^*/dT)_R(T_0 - T_1), \quad (R \text{ indicates a reference temp. near } (T_0 + T_1)/2)$$

$$H_L = s^* H_S + \varrho L C_H V_1(q^*(T_1) - q_1), \tag{11.1}$$

$$s^* = (L/c_p)(dq^*/dT)_R \; (= 0.7 \text{ at } 273 \text{ K}, 3.3 \text{ at } 300 \text{ K})$$

This equation expresses latent heat flux in terms of sensible heat flux and the saturation deficit at the measurement level. It is immediately apparent that the Bowen ratio H_S/H_L must be at most s^{*-1} over a saturated surface, and that it drops as the relative humidity of the overlying air decreases. At higher temperatures, latent heat fluxes tend to become more dominant. For an ideal surface, (11.1), together with energy balance

$$R_N - H_G = H_S + H_L$$

can be solved for H_L:

$$H_L = LE_P = \Gamma(R_N - H_G) + (1 - \Gamma)\, \varrho L_v C_H V_1 (q^*(T_1) - q_1) \tag{11.2}$$

$$\Gamma = s^* /(s^* + 1) \; (= 0.4 \text{ at } 273 \text{ K}, 0.77 \text{ at } 300 \text{ K})$$

The corresponding evaporation rate E_P is called the **potential evaporation**, and is the maximum possible evaporation rate given the surface characteristics and the atmospheric state at the measurement height. If the surface is not saturated, the evaporation rate will be less than E_P. The figure on the next page shows H_L vs. the net surface energy influx $R_N - H_G$ for T_1 = 293 K and RH_1 = 57%, at a height of z_1 = 10 m, with a geostrophic wind speed of 10 m s^{-1}, assuming a range of surface roughness. Especially over rough surfaces (forest), H_L often exceeds R_N - H_G, so the sensible heat flux must be negative by up to 100 W m^{-2}. The Bowen ratio is quite small (0.2 or less) for all the saturated surfaces shown in this figure.

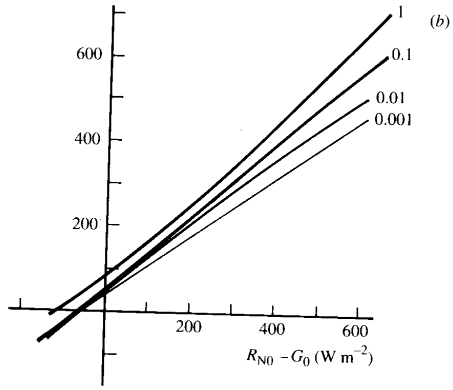


Fig. 5.6 Potential evaporation for different wet surfaces calculated from Eq. 5.26. In (*a*) neutral conditions have been assumed, and in (*b*) the full stability correction in r_{aV} is included (see Eqs. 3.47 and 3.57). Note how the effects of thermal stability tend to reduce the direct influence of aerodynamic roughness. Values of z_0 are as follows: 0.001 m, lake; 0.01 m, grass; 0.1 m, scrub; 1 m, forest. Further details of the calculations can be found in Webb (1975).

Evaporation from dry vegetation

We consider a fully vegetated surface with a single effective surface temperature and humidity (a 'single-layer canopy'). The sensible heat flux originates at the leaf surfaces, whose temperature is T_0. The latent heat flux is driven by evaporation of liquid water out of the intercellular spaces within the leaves through the stomata, which are channels from the leaf interior to its surface. The evaporation is proportional to the humidity difference between the saturated inside of the stomata and the ambient air next to the leaves. The constant of proportionality is called the **stomatal resistance** (units of inverse velocity)

$$r_{st} = \rho(q^*(T_0) - q_0)/E \qquad (11.3)$$

Plants regulates transport of water vapor and other gasses through the stomata to maintain an optimal internal environment, largely shutting down the stomata when moisture-stressed. Hence r_{st} depends not only on the vegetation type, but also soil moisture, temperature, etc. Table 5.1 of Garratt shows measured r_{st}, which vary from 30 -300 s m^{-1}.

By analogy, we can define an **aerodynamic resistance**

$$r_a = (C_H V_1)^{-1} = \rho(q_0 - q_1)/E \qquad (11.4)$$

Typical values of r_a are 100 s m^{-1}, decreasing in high wind or highly convective conditions. This is comparable to the stomatal resistance. Working in terms of aerodynamic resistance in place of C_H is convenient in this context, as we shall see next, because these resistances add:

$$r_{st} + r_a = \rho(q^*(T_0) - q_0)/E + \rho(q_0 - q_1)/E = \rho(q^*(T_0) - q_1)/E, \qquad (11.5)$$

i. e. E is identical to the evaporation rate over an equivalent saturated surface with aerodynamic resistance $r_{st} + r_a$. The same manipulations that led to (11.1) and (11.2) now lead to:

$$H_S = \varrho c_p(T_0 - T_1)/r_a$$

$$H_L = LE = \varrho(q^*(T_0) - q_1)/(r_{st} + r_a) = \{s^*H_S + \varrho L(q^*(T_1) - q_1)\}\{r_a/(r_{st} + r_a)\}$$

$$H_L = \Gamma^*(R_N - H_G) + (1 - \Gamma^*)\rho L(q^*(T_1) - q_1)/ (r_{st} + r_a) \,, \tag{11.6}$$

where $\Gamma^* = s^* /(s^* + 1 + r_{st}/r_a)$

This is the **Penman-Monteith** relationship. Comparing (11.6) to (11.2), we find that $\Gamma^* < \Gamma$, so the heat flux will be partitioned more into sensible heating if stomatal resistance is high, winds are high, or the BL is unstable. The effect is magnified at cold temperatures where s^* is small. The ratio of H_L to the saturated latent heat flux (11.2) given the same energy influx R_N - H_G is

$$\frac{H_L}{H_{L,sat}} = \frac{1}{1+(1-\Gamma)r_{st}/r_a} \tag{11.7}$$

Calculations of this ratio for neutral conditions, a 10 m s^{-1} geostrophic wind speed, and various surface roughnesses are shown in the figure below. For short grass, the surface transfer coefficient is low, so the aerodynamic resistance is high and stomatal resistance does not play a crucial role at high temperatures (though at low temperatures it cuts off a larger fraction of the latent heat flux). For forests, stomatal resistance is very important due to the high surface roughness (low aerodynamic resistance).

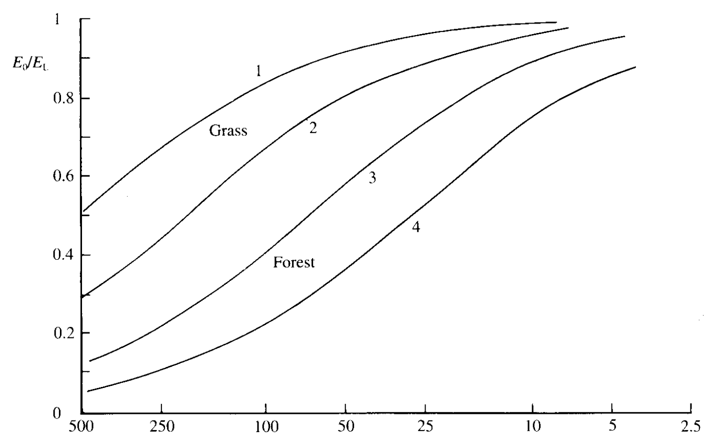


Fig. 5.8 Variations of E_0/E_L (Eq. 5.37) with surface resistance. Values of r_{aV} have been calculated for neutral conditions, with $z_q = z_0/7.4$. For short grass ($z_0 = 0.0025$ m): curve 1, $T = 303$ K; curve 2, $T = 278$ K. For forest ($z_0 = 0.75$ m): curve 3, $T = 303$ K; curve 4, $T = 278$ K.

Soil moisture

If the surface is partly or wholly unvegetated, the evaporation rate depends on the available soil moisture. Soil moisture is also important because it modulates the thermal conductivity and hence the ground heat flux, and affects the surface albedo as well as transpiration by surface vegetation. For instance, Idso et al. (1975) found that for a given soil, albedo varied from 0.14 when the soil was moist to 0.31 when it was completely dry at the surface.

Rooted plants also rely on soil moisture and also are a major sink of soil moisture in the rooting zone during their growing season. The equations for the vertical movement of soil moisture discussed below need to include this sink.

If the soil-surface relative humidity RH_0 is known, then the evaporation is

$$E = \rho(RH_0\, q^*(T_0) - q_1)/r_a. \tag{11.8}$$

Note that net evaporation ceases when the mixing ratio at the surface drops below the mixing ratio at the measurement height, which does not require the soil to be completely dry. Soil moisture can be expressed as a volumetric moisture content η (unitless), which does not exceed a saturated value η_s, usually around 0.4. When the soil is saturated, moisture can easily flow through it, but not all pore spaces are water-filled. As the soil becomes less saturated, water is increasingly bound to the soil by adsorption (chemicals) and surface tension.The movement of water through the soil is down the gradient of a combined gravitational potential gz (here $z < 0$ below the surface) plus a moisture potential $g\psi(\eta)$. The moisture potential is always negative, and becomes much more so as the soil dries out and its remaining water is tightly bound. Note ψ has units of height. The downward flux of water is

$$F_w = -\rho_w K(\eta)\partial(\psi + z)/\partial z, \text{ (Darcy's law)} \tag{11.9}$$

where $K(\eta)$ is a hydraulic conductivity (units of m s^{-1}), which is a very rapidly increasing function of soil moisture. Neglecting sinks (e.g. roots), conservation of soil moisture requires

$$\rho_w \partial\eta/\partial t = -\partial F_w/\partial z \tag{11.10}$$

The surface relative humidity is

$$\mathrm{RH}_0 = \exp(-g\psi|_{z=0}/R_v T_0) \tag{11.11}$$

i. e. the more tightly bound the surface moisture is to the soil, the less it is free to evaporate. Empirical forms for ψ and K as functions of η have been fitted to field data for various soils:

$$\psi = \psi_s(\eta/\eta_s)^{-b} \tag{11.12}$$

$$K = K_s(\eta/\eta_s)^{2b+3} \tag{11.13}$$

where ψ_s and K_s are saturation values, depending on the soil, and the exponent b is 4-12. For b = 5, halving the soil moisture increases the moisture potential by a factor of 32 and decreases the hydraulic conductivity by a factor of 4000! Because these quantities are so strongly dependent on η, one can define a critical surface soil moisture, the wilting point η_w, above which the surface relative humidity RH_0 is larger than 99%, and below which it rapidly drops. The wilting point can be calculated as the η below which the hydraulic suction $-\psi$ exceeds 150 m.

Land surface schemes

All regional and global models have land surface parameterizations that handle the many land surface and vegetation types, including forest canopies, as well as soil moisture, and determine the surface turbulent and radiative fluxes from the local atmospheric conditions and their prior history. These would take an entire course to do justice to, so we just note some popular choices:

For WRF and NCEP's GFS forecast model: Noah (Ek et al. 2003 *JGR*)
http://www.ral.ucar.edu/research/land/technology/lsm.php

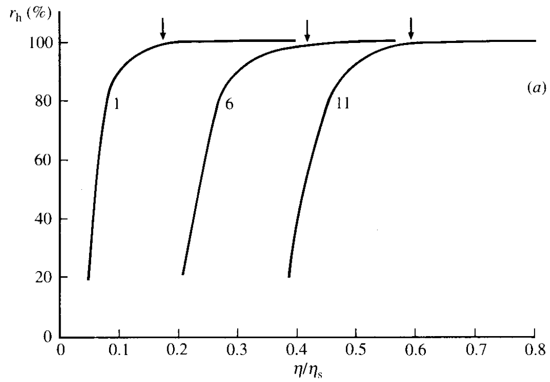
For CESM climate model: Community Land Model (CLM) (Lawrence et al. 2011 *JAMES*):
http://www.cesm.ucar.edu/models/clm/

Each of these is a community-developed and supported model that is undergoing constant evolution and refinement.

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 ($m^3\,m^{-3}$)	ψ_s (m)	$K_{\eta s}$ ($10^{-6}\,m\,s^{-1}$)	b	η_w ($m^3\,m^{-3}$)
1. sand	0.395	− 0.121	176	4.05	0.0677
2. loamy sand	0.410	− 0.090	156.3	4.38	0.075
3. sandy loam	0.435	− 0.218	34.1	4.90	0.1142
4. silt loam	0.485	− 0.786	7.2	5.30	0.1794
5. loam	0.451	− 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	− 0.356	1.7	7.75	0.2181
8. clay loam	0.476	− 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	− 0.490	1.0	10.40	0.2832
11. clay	0.482	− 0.405	1.3	11.40	0.2864



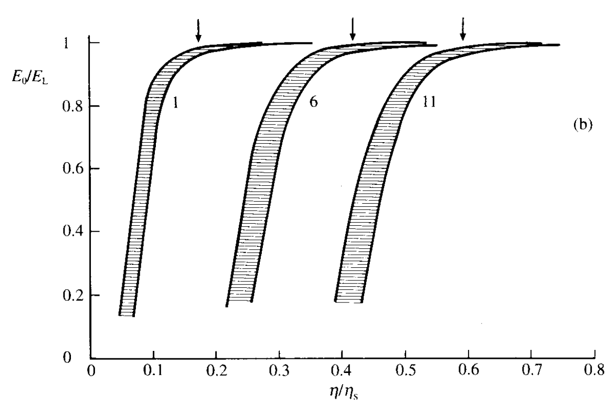


Fig. 5.9 (*a*) Relative humidity r_h as a function of relative soil moisture content η/η_s, based on Eq. 5.49 and data in Table A9 for soil types 1 (sand), 6 (loam) and 11 (clay). Calculations are for a temperature T_0 of 303 K. The vertical arrows indicate the wilting points. Note that combining Eqs. 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 $q = 0.005$).