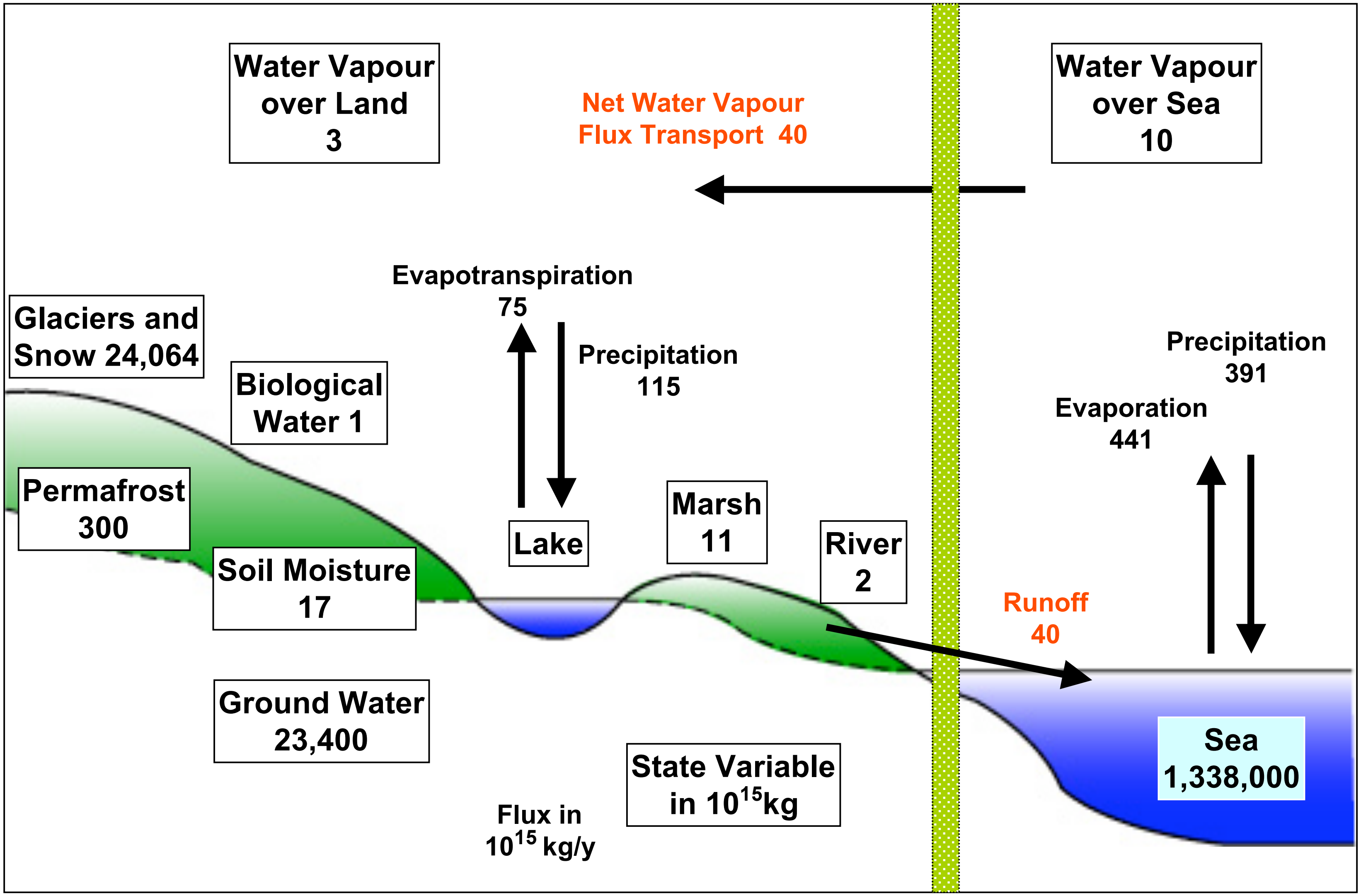
Global Water Cycle

Surface (ocean and land): source of water vapor to the atmosphere



NCAR

Lecture 11. Surface Evaporation (Garratt 5.3)

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

$$\boxed{\begin{aligned} H_S &= \rho c_p C_H V_1 (T_0 - T_1), \\ H_L &= \rho L C_H V_1 (q_0 - q_1), \end{aligned}} \quad \text{where} \quad C_H = \mathrm{fn}(V_1, z_1, z_0, z_T, L)$$

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

$$\boxed{\begin{aligned} H_S &= \rho c_p C_H V_1(T_0 - T_1), \\ H_L &= \rho L C_H V_1(q_0 - q_1), \end{aligned}} \quad \text{where} \quad C_H = \text{fn}(V_1, z_1, z_0, z_T, L)$$

We can eliminate $\cancel{T_0}$ using a linearized version of the ~~Clausius-Clapeyron~~ equations:

$$\boxed{q_0 - q^*(T_1) = (dq^*/dT)_R(T_0 - T_1),} \quad R \text{ indicates a value at a reference temperature,}$$

that ideally should be close to $(T_0 + T_1)/2$

$$\boxed{H_L = s^* H_S + \rho L C_H V_1(q^*(T_1) - q_1),} \quad \boxed{s^* = (L/c_p)(dq^*/dT)_R} (= 0.7 \text{ at } 273 \text{ K}, 3.3 \text{ at } 300 \text{ K}) \qquad (1)$$

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, es. At higher temperatures, latent heat fluxes tend to become more dominant. For an ideal surface, (1), together with energy balance

$$\boxed{R_N - H_G = H_S + H_L}$$

can be solved for H_L:

$$\boxed{H_L = LE_P = \Gamma(R_N - H_G) + (1 - \Gamma)\rho L C_H V_1 (q^*(T_1) - q_1)} \quad (2)$$

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

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.

Garratt

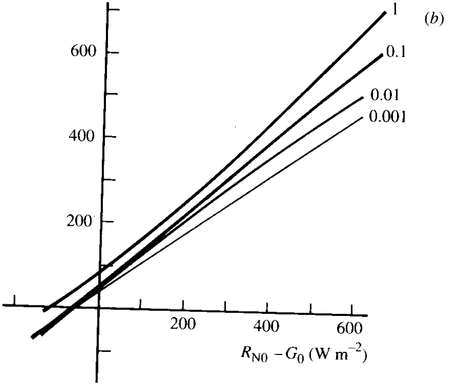
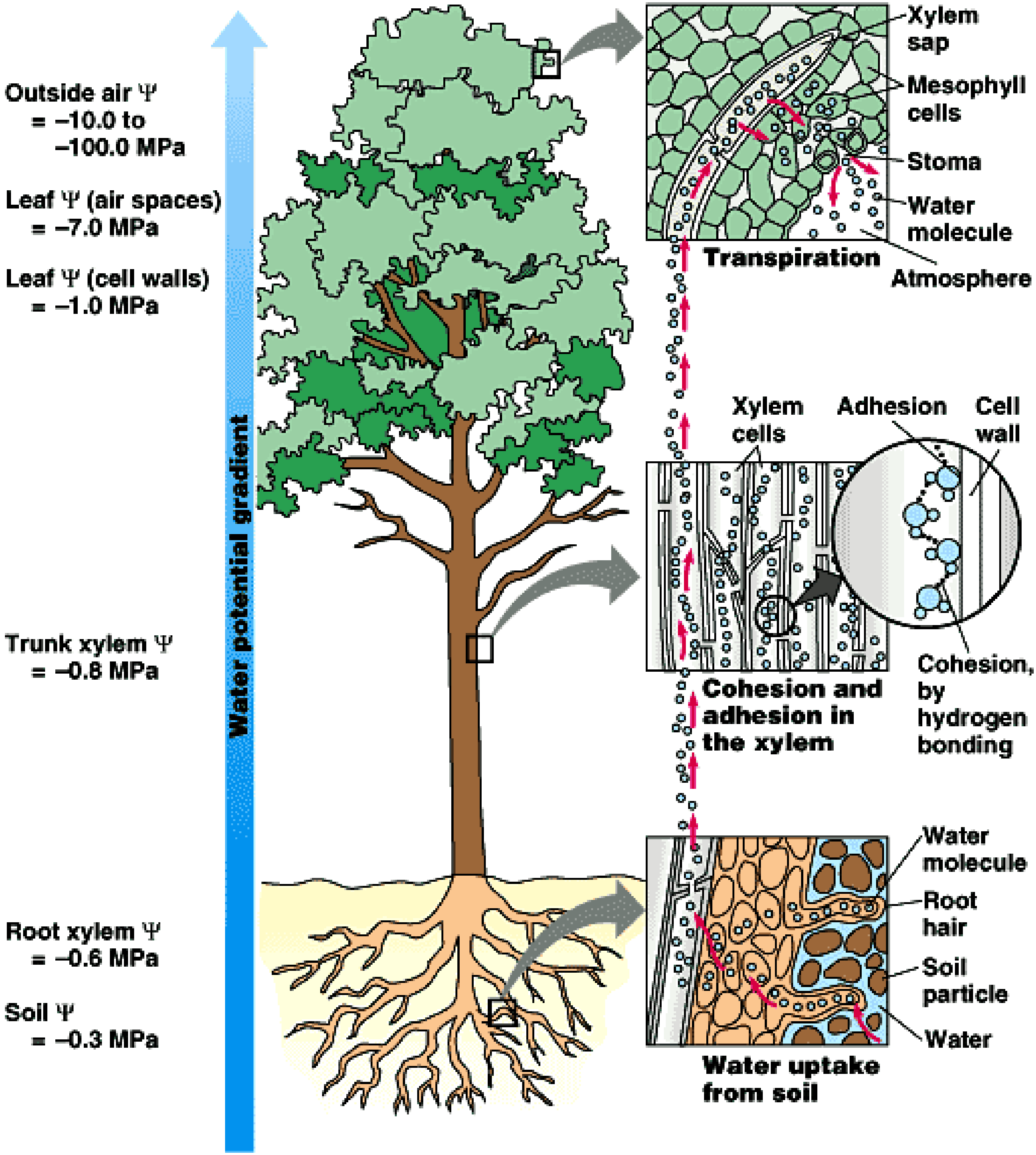
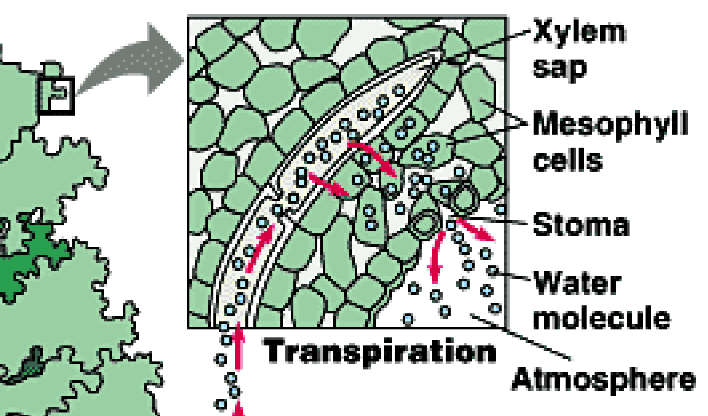


Fig. 5.6 Potential evaporation for different wet surfaces calculated from Eq. 5.26. In (a) neutral conditions have been assumed, and in (b) the full stability correction in r_{aV} is included (see Eqs. 3.47 and 3.57). Note how the effects of thermal stability tend to reduce the direct influence of aerodynamic 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).



Outside air Ψ
= –10.0 to
–100.0 MPa
Leaf Ψ (air spaces)
= –7.0 MPa
Leaf Ψ (cell walls)
= –1.0 MPa
Trunk xylem Ψ
= –0.8 MPa
Root xylem Ψ
= –0.6 MPa
Soil Ψ
= –0.3 MPa
Water potential gradient
Xylem sap
Mesophyll cells
Stoma
Water molecule
Atmosphere
Transpiration
Xylem cells
Adhesion
Cell wall
Cohesion, by hydrogen bonding
Cohesion and adhesion in the xylem
Water molecule
Root hair
Soil particle
Water
Water uptake from soil


Xylem sap
Mesophyll cells
Stoma
Water molecule
Transpiration
Atmosphere

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 is 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)

$$\boxed{r_{st} = \rho(q^*(T_0) - q_0)/E} \tag{3}$$

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 varies form 30 -300 s m^{-1}.

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

$$\boxed{r_a = (C_H V_1)^{-1} = \rho(q_0 - q_1)/E} \tag{4}$$

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

$$\boxed{r_a = (C_H V_1)^{-1} = \rho(q_0 - q_1)/E} \qquad (4)$$

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:

$$\boxed{r_{st} + r_a = \rho(q^*(T_0) - q_0)/E + \rho(q_0 - q_1)/E = \rho(q^*(T_0) - q_1)/E,} \qquad (5)$$

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 (1) and (2) now lead to:

$$\boxed{\begin{aligned} &H_S = \rho c_p(T_0 - T_1)/r_a \\ &H_L = LE = \rho(q^*(T_0) - q_1)/(r_{st} + r_a) = \{s^* H_S + \rho L(q^*(T_1) - q_1)\}\{r_a/(r_{st} + r_a)\} \\ &H_L = \Gamma^*(R_N - H_G) + (1 - \Gamma^*)\rho L(q^*(T_1) - q_1)/(r_{st} + r_a) \quad , \end{aligned}} \qquad (6)$$

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

This is the **Penman-Monteith** relationship. Comparing (6) to (2), we find that $\Gamma^* < \Gamma$, so the heat flux will be partitioned more into sensible heating, especially 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 (2) given the same energy influx R_N - H_G is

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

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

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).

Garratt

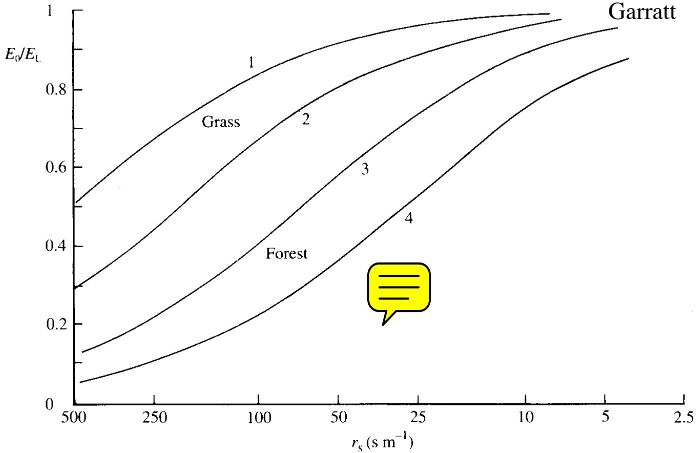


Fig. 5.8 Variations of E_0/E_L (Eq. 5.37) with surface resistance. Values of r_{aV} have been calculated for neutral conditions, with $z_q = z_0/7.4$. For short grass ($z_0 = 0.0025$ m): curve 1, $T = 303$ K; curve 2, $T = 278$ K. For forest ($z_0 = 0.75$ m): curve 3, $T = 303$ K; curve 4, $T = 278$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

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

$$\boxed{E = \rho(\mathrm{RH}_0 q^*(T_0) - q_1)/r_a \ .}$$

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 we take z as depth 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

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

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

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

The surface relative humidity is

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

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:

$$\boxed{\begin{aligned} \psi &= \psi_s(\eta/\eta_s)^{-b} \\ K &= K_s(\eta/\eta_s)^{2b+3} \end{aligned}}$$

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.

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

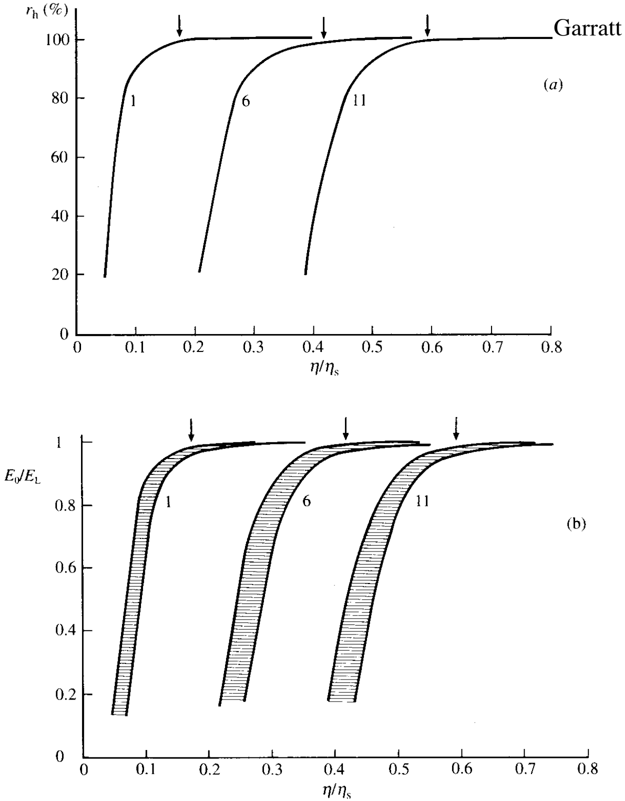


Fig. 5.9 (a) Relative humidity r_h as a function of relative soil moisture content η/η_s, based on Eq. 5.49 and data in Table A9 for soil types 1 (sand), 6 (loam) and 11 (clay). Calculations are for a temperature T_0 of 303 K. The vertical arrows indicate the wilting points. Note that combining Eqs.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

Parameterization of surface evaporation in large-scale models

In practice, simplified formulations of soil moisture and transpiration are used in most models. We will defer most of these until later. However, the MM5 formulation of surface evaporation is particularly simplified. It is

$$\boxed{E = \rho L C_H V_1 M(q^*(T_0) - q_1),}$$

i. e. the standard formula for evaporation off a *saturated* surface at the ground temperature T_0 (calculated by the model) multiplied by a moisture availability factor M between 0 and 1 that is assumed to depend only on the surface type. This formulation avoids the need to initialize soil moisture, but is tantamount to assuming a surface resistance that is proportional to the aerodynamic resistance, with

$$r_s/r_a = (1 - M)/M$$

While this type of formulation can be tuned to give reasonable results on an annually averaged basis, it is likely to be in error by a factor of two or more in individual situations, because r_s and r_a are both subject to large and independent fluctuations. More sophisticated schemes explicitly prognose soil moisture (often using relaxation to specified values deep within the soil to control fluctuations) and vegetation characteristics and determine the evaporation from these.

Landuse Integer Identification	Landuse Description	Albedo (%)		Moisture Avail. (%)		Emissivity (% at 9 μm)		Roughness Length (cm)		Thermal Inertia (cal cm^{-2} K^{-1} $s^{-1/2}$)	
		Sum	Win	Sum	Win	Sum	Win	Sum	Win	Sum	Win
1	Urban land	18	18	5	10	88	88	50	50	0.03	0.03
2	Agriculture	17	23	30	60	92	92	15	5	0.04	0.04
3	Range-grassland	19	23	15	30	92	92	12	10	0.03	0.04
4	Deciduous forest	16	17	30	60	93	93	50	50	0.04	0.05
5	Coniferous forest	12	12	30	60	95	95	50	50	0.04	0.05
6	Mixed forest and wet land	14	14	35	70	95	95	40	40	0.05	0.06
7	Water	8	8	100	100	98	98	.0001	.0001	0.06	0.06
8	Marsh or wet land	14	14	50	75	95	95	20	20	0.06	0.06
9	Desert	25	25	2	5	85	85	10	10	0.02	0.02
10	Tundra	15	70	50	90	92	92	10	10	0.05	0.05
11	Permanent ice	55	70	95	95	95	95	5	5	0.05	0.05
12	Tropical or sub-tropical forest	12	12	50	50	95	95	50	50	0.05	0.05
13	Savannah	20	20	15	15	92	92	15	15	0.03	0.03

MM5 surface types and their characteristics (Appendix 4 of MM5 manual)

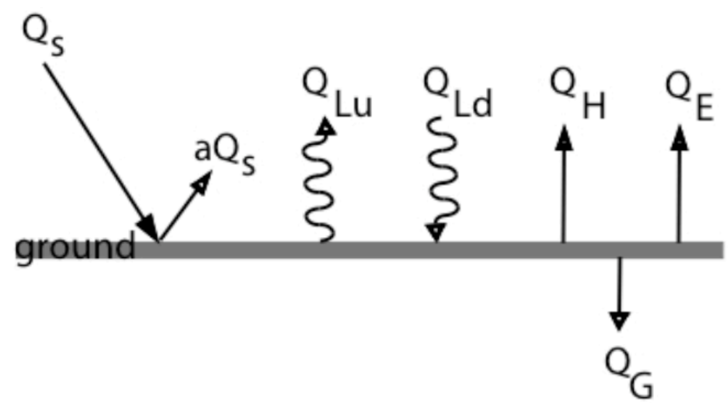
Why Do We Need Land Surface Models?

- Need to account for subgrid-scale fluxes
- The lower boundary is the only physical boundary for atmospheric models
- LSM becomes increasingly important:
  - More complex PBL schemes are sensitive to surface fluxes and cloud/cumulus schemes are sensitive to the PBL structures
  - NWP models increase their grid-spacing (1-km and sub 1-km). Need to capture mesoscale circulations forced by surface variability in albedo, soil moisture/temperature, landuse, and snow
- Not a simple task: tremendous land surface variability and complex land surface/hydrology processes
- Initialization of soil moisture/temperature is a challenge

Land-surface model (LSM) development chronology

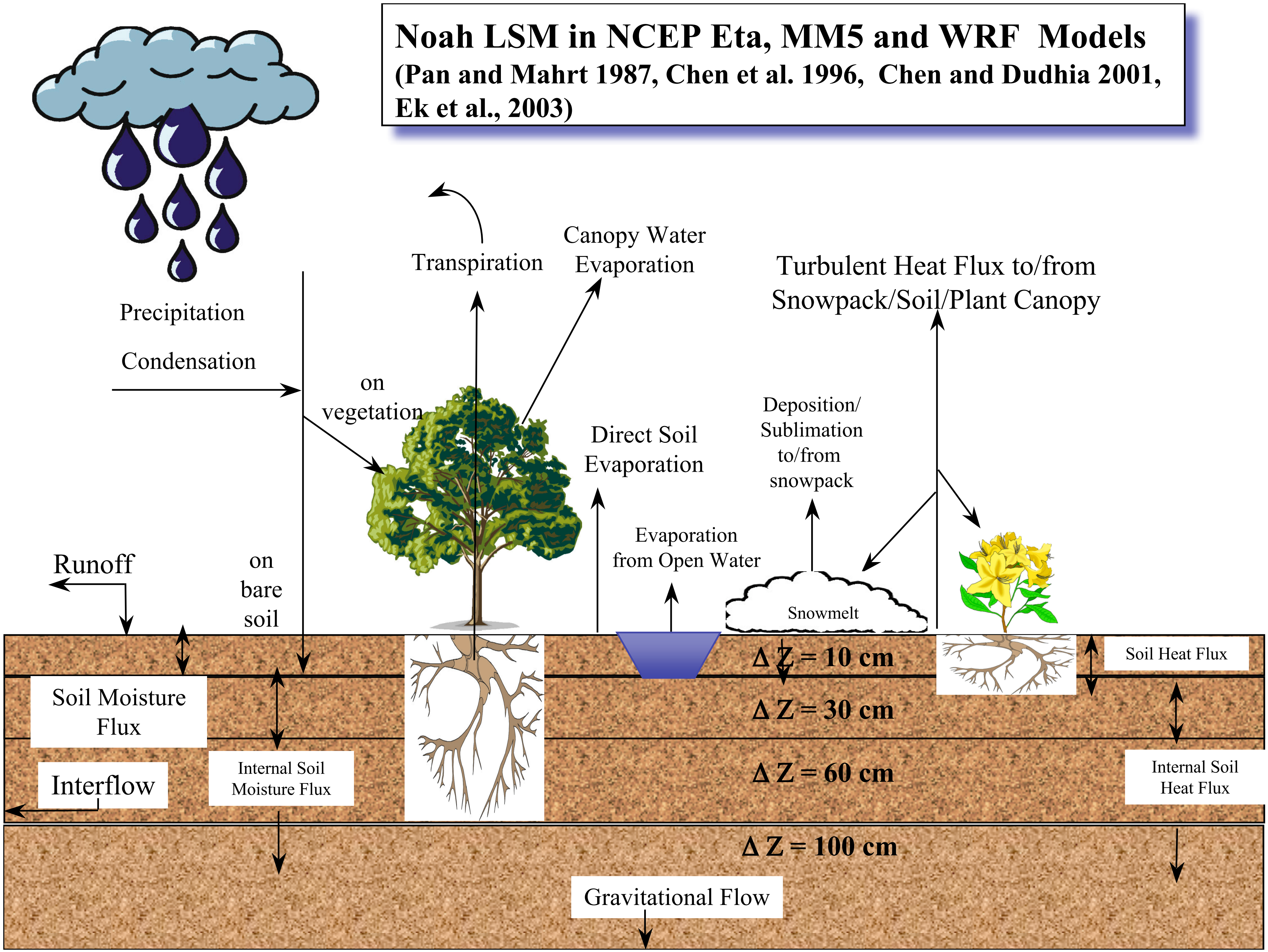
- Gen-0 (prior to 60s): lack of land-surface processes (prescribed diurnal cycle of surface temperature)
- Gen-1a (mid 60s): surface model with time-fixed soil moisture
- Gen-1b (late 60s): Bucket Model (Manabe 1969): time- and space-varying soil moisture
- Gen-2 (70s): Big-leaf model (Deardorff 1978): explicit vegetation treatment; a major milestone
- Gen-3 (late 80s): development of more sophisticated models including hydrological, biophysical, biochemical, ecological processes (e.g., BATS, SiB, NCARLSM, Century)
- mid 90s: implementation of advanced LSMs at major operational numerical weather prediction (NWP) centers

An LSM must provide 4 quantities to parent atmospheric model

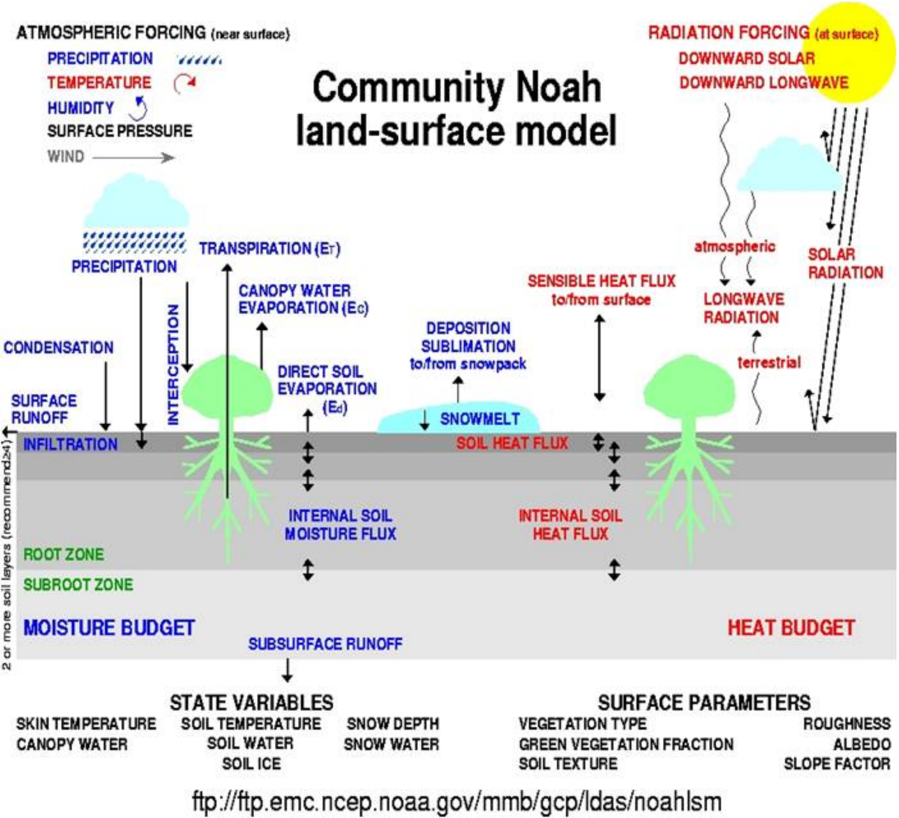


- surface sensible heat flux Q_H
- surface latent heat flux Q_E
- upward longwave radiation Q_{Lu}
  - Alternatively: skin temperature and sfc emissivity
- upward (reflected) shortwave radiation aQ_S
  - Alternatively: surface albedo, including snow effect

NCAR



Noah LSM in NCEP Eta, MM5 and WRF Models
(Pan and Mahrt 1987, Chen et al. 1996, Chen and Dudhia 2001, Ek et al., 2003)
Precipitation
Condensation
on vegetation
on bare soil
Runoff
Transpiration
Canopy Water Evaporation
Direct Soil Evaporation
Evaporation from Open Water
Deposition/ Sublimation to/from snowpack
Turbulent Heat Flux to/from Snowpack/Soil/Plant Canopy
Snowmelt
Soil Heat Flux
Soil Moisture Flux
Interflow
Internal Soil Moisture Flux
Internal Soil Heat Flux
Δ Z = 10 cm
Δ Z = 30 cm
Δ Z = 60 cm
Δ Z = 100 cm
Gravitational Flow



ATMOSPHERIC FORCING (near surface)
PRECIPITATION
TEMPERATURE
HUMIDITY
SURFACE PRESSURE
WIND
Community Noah land-surface model
RADIATION FORCING (at surface)
DOWNWARD SOLAR
DOWNWARD LONGWAVE
PRECIPITATION
TRANSPIRATION (ET)
CANOPY WATER EVAPORATION (Ec)
INTERCEPTION
CONDENSATION
SURFACE RUNOFF
DIRECT SOIL EVAPORATION (Ed)
DEPOSITION SUBLIMATION to/from snowpack
SNOWMELT
SENSIBLE HEAT FLUX to/from surface
atmospheric
LONGWAVE RADIATION
terrestrial
SOLAR RADIATION
INFILTRATION
SOIL HEAT FLUX
INTERNAL SOIL MOISTURE FLUX
INTERNAL SOIL HEAT FLUX
ROOT ZONE
SUBROOT ZONE
2 or more soil layers (recommend≥4)
MOISTURE BUDGET
HEAT BUDGET
SUBSURFACE RUNOFF
STATE VARIABLES
SKIN TEMPERATURE
CANOPY WATER
SOIL TEMPERATURE
SOIL WATER
SOIL ICE
SNOW DEPTH
SNOW WATER
SURFACE PARAMETERS
VEGETATION TYPE
GREEN VEGETATION FRACTION
SOIL TEXTURE
ROUGHNESS
ALBEDO
SLOPE FACTOR
ftp://ftp.emc.ncep.noaa.gov/mmb/gcp/ldas/noahlsm