Evapotranspiration

- **Evaporation** (E): In general, the change of state from liquid to gas
  - Here, liquid water on surfaces or in the very thin surface layer of the soil that **evaporates directly to the atmosphere**

- **Transpiration** (T): vapour loss from stomata in plant leaves

- **Evapotranspiration** (ET): net transfer (loss) of water vapour from wet surfaces (rivers, lakes, soil) & vegetation into the atmosphere…
  - each process **difficult to measure separately**… often combined as net evapotranspiration

Free-water loss

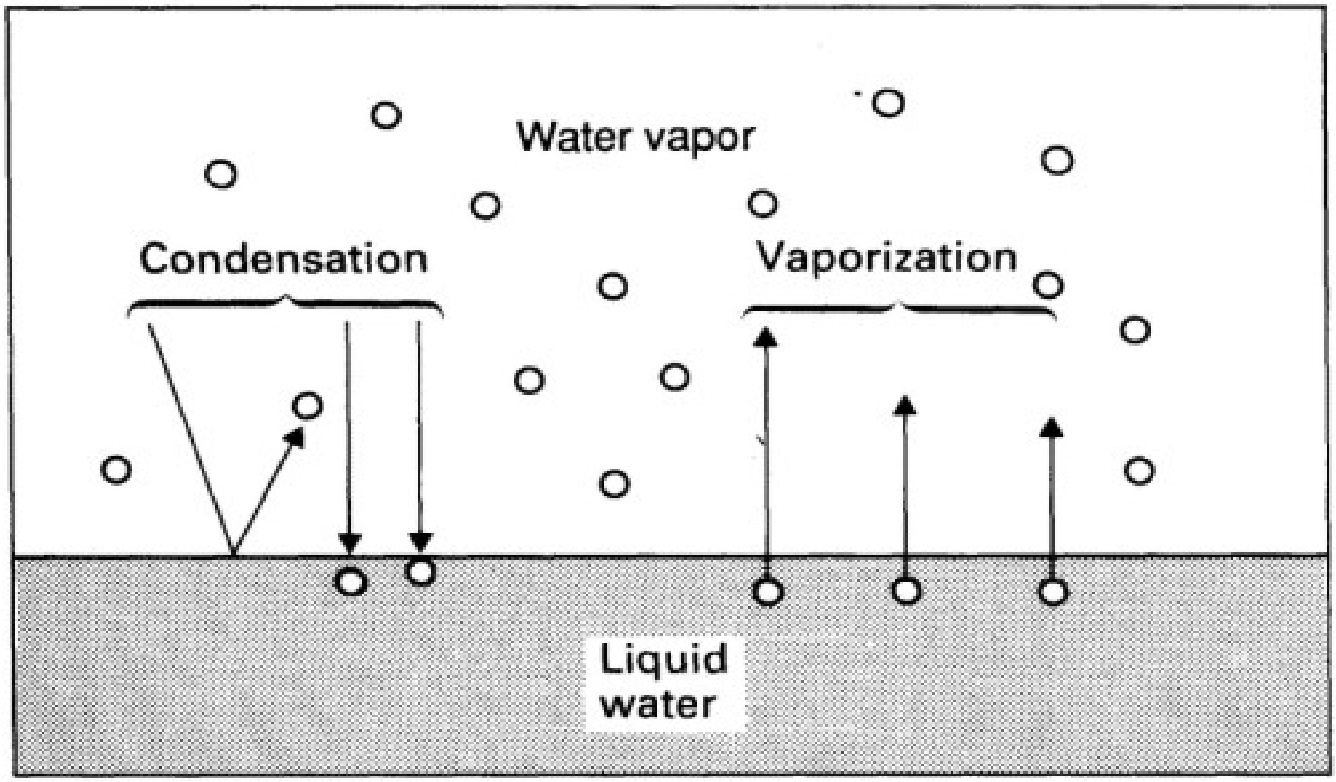


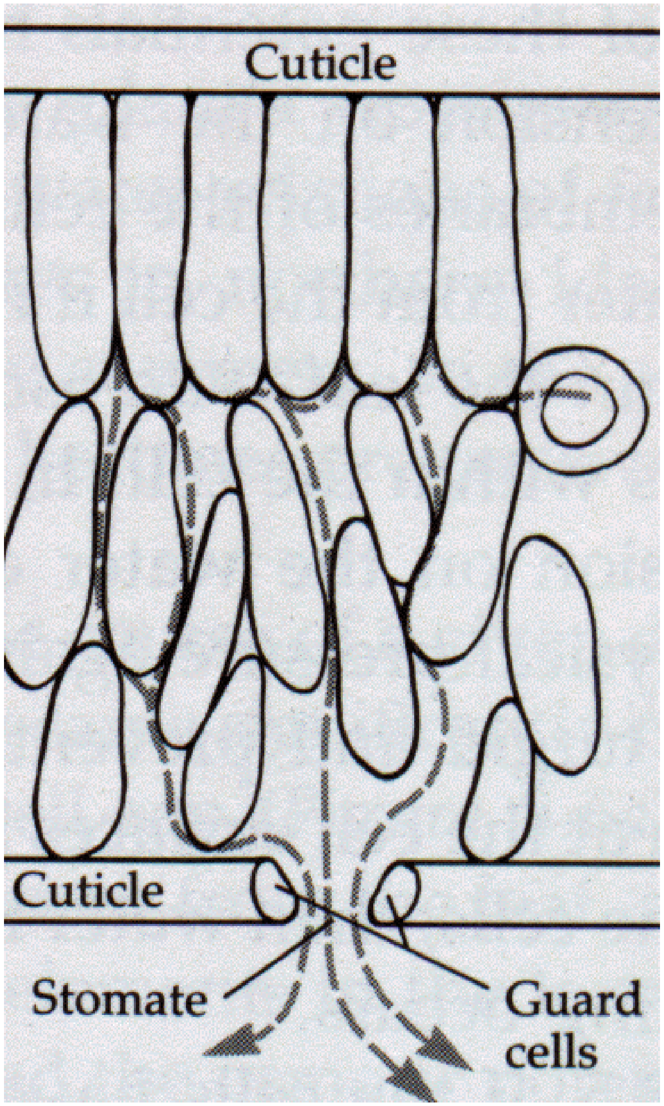
FIGURE 4.2.1 Molecular exchange between liquid water and water vapor. Not all the molecules hitting the surface are captured, but some condense at a rate which is proportional to the vapor pressure of the moist air: molecules with enough energy vaporize at a rate determined by the surface temperature.

Maidment (1993)

Water movement through the soil-plant-atmosphere continuum

- Soil water that can be freed from the soil can proceed to the atmosphere in **two ways**:
- **Evaporation** - Water in the soil evaporates **directly** into the atmosphere. Evaporation only affects the **thin surface layer** of soils, as the resistance to liquid water movement in soils is high
- **Transpiration** - Plants provide an ideal conduit for the movement of water between soils and the atmosphere. Roots grow deep into the soil and can tap into **water reserves far from the surface**, providing a pathway between the deeper soil and the atmosphere

Water movement through the soil-plant-atmosphere continuum



- The movement of water from the soil through plant and into the atmosphere is controlled by **stomata**, tiny holes on the back of leaves
- The **atmosphere** is usually **drier** than the **air inside the stomata**, thus there exists a **water potential gradient** (the potential in the outside air is more negative) causing the water move from the stomata into the atmosphere

Water movement through the soil-plant-atmosphere continuum

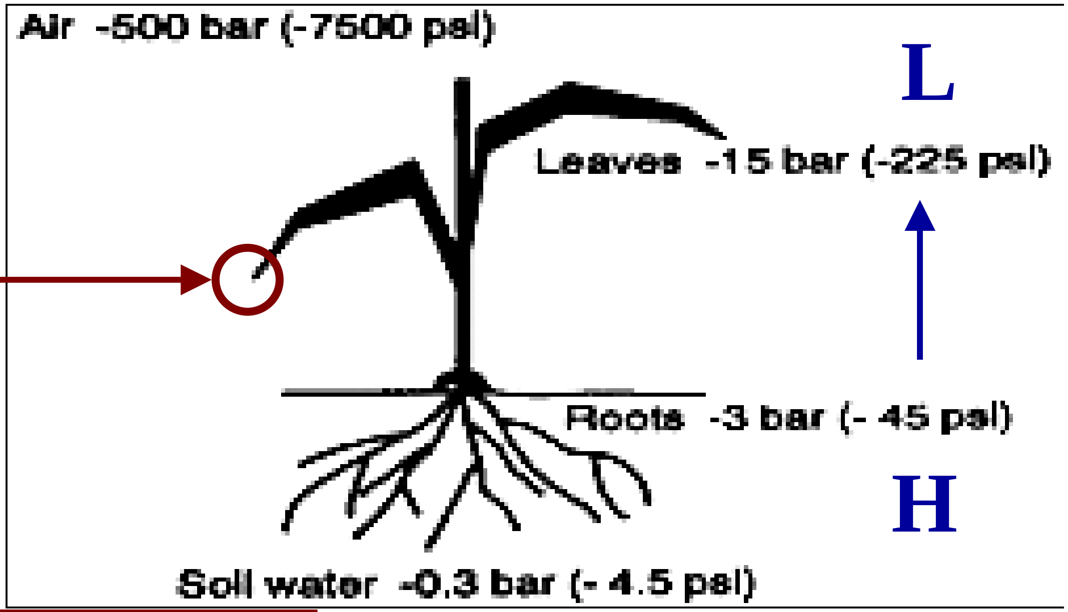
- The negative water potential in the atmosphere is transferred to a **continuous column of liquid water** that begins in the root and ends in the leaf
- The tissue that the water passes through is called **xylem**, which provides an uninterrupted pathway for water movement
  - The tension is conducted out through the roots, and through contact between roots and soil, to the water adhering to soil particles
- This water column **must be continuous**. Any air gaps in the system will relieve the tension and stop the movement of water. Root surfaces must be in direct contact with the soil water film

Water movement through the soil-plant-atmosphere continuum

- The actual **rate of flow** of water up through the plant, and thus from the soil to the atmosphere is a **function** of the **differences in water potential** between these two ends of the gradient and the **resistance to the flow**
- Resistance within the plant results mainly from **friction** between water and the walls of the xylem elements through which it passes
- The **force of gravity** also works against the rate of water movement up the stem
- During times of **water stress**, the guard cells lose water, reducing the turgor of the cells. As the guard cell loses turgor, the **stomata will close**, to further reduce the loss of water

Transpiration loss

- a **major component** of vapor exchange at soil – atmosphere interface…



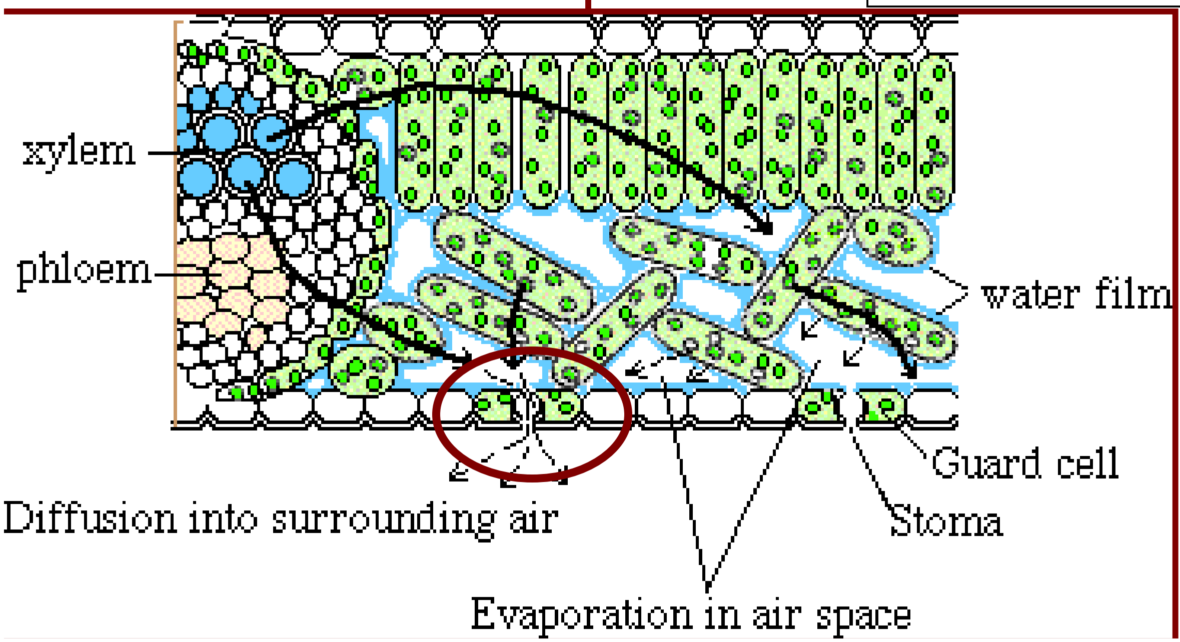


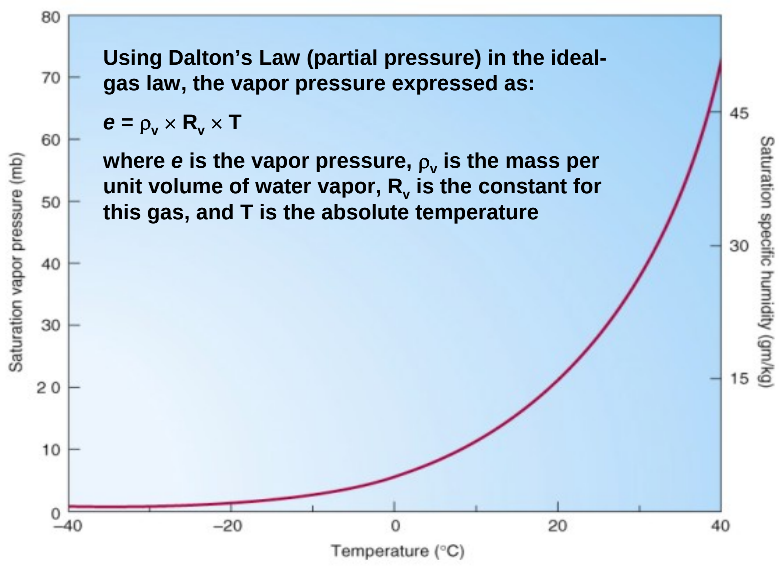
Fig. 1: vapor pressure (e_a) at soil – plant – atmosphere interface

Controlling factors for ET

1. **Water**:
   - open bodies, intercepted, soil, plants

2. **Energy**:
   - major source is <u>short-wave</u> solar radiation
   - <u>long-wave</u> (sensible heats surfaces) &
   - <u>latent heat</u> (exchanged within air masses)

3. **Vapor pressure (humidity)**:
   - Difference between atmosphere & water source
   - pressure gradient controls rates of movement of H_2O molecules from moist surfaces to atm.
   - recall, $e_a \leq e^*$ or $e_a \leq e_s$
   - cannot exceed RH =100%



Controlling factors for ET

4. **wind**:
   - turbulent airflow above moist surfaces removes saturated air replacing it with unsaturated air (↓e_a)

5. **vegetation**:
   - transpiration is a product of photosynthesis
     - uses soil moisture
   - rates controlled by e_a
   - also includes E_i losses from plant canopies
   - E from bare soil may actually » T from veg… why?

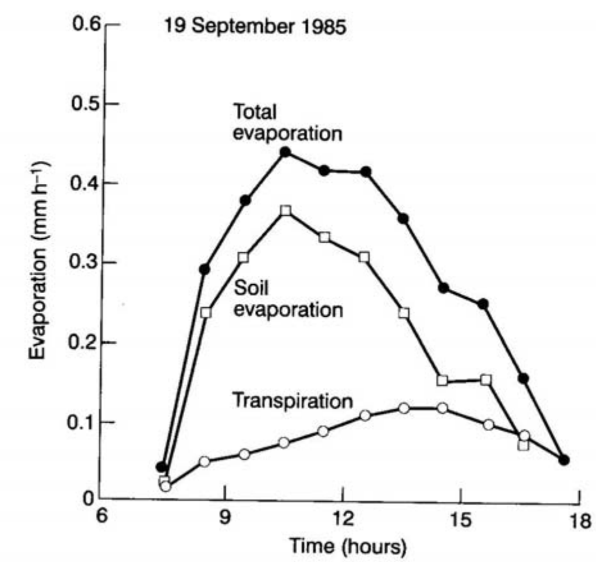


Figure 4.11 Direct soil evaporation may, after rainfall, account for twice the water loss resulting from transpiration (from Wallace *et al.*, 1989).

Physics of evaporation

- Evaporation is a **diffusive** process …a function of exchange across ‘gradients’
  - **energy gradient**: latent- & sensible-heat exchange [E L^{-2} T^{-1}]
  - **vapor pressure gradient**: difference between $e_{surface}$ & e_{air} (essentially a ‘mass’ gradient in [M L^{-1} T^{-2}])
  - **windspeed** (v_a) & **turbulent kinetic energy** control vertical transport (removal) of H_2O vapor

Wind & turbulent energy effects

Described by **Fick's 1st Law**: $E = K_e \cdot v_a \cdot (e_s - e_a)$

- K_e is a coefficient that describes the efficiency of vertical transport of water vapor via turbulent eddies

- ρ_a = air density
- ρ_w = water density
- P = atmospheric pressure
- z_m is height windspeed V_a & e_a measured
- z_d is zero-plane displacement height
- z_0 is the aerodynamic roughness height

$$K_e \equiv \frac{0.622\,\rho_a}{P \cdot \rho_w} \cdot \left[6.25 \left(\ln \left[\frac{z_m - z_d}{z_0} \right] \right)^2 \right]^{-1}$$

Latent heat exchange (LE)

- LE is 'lost' during **vaporization** (λ_v) & causes a **reduction in T_s** (i.e., cooling of surface) …. Example?
  - if we measure ΔLE, we know amount of energy avail. for evaporation

- LE = $\rho_w \cdot \lambda_v \cdot E \quad = \rho_w \cdot \lambda_v \cdot K_e \cdot v_a (e_s - e_a)$

  - λ_v **latent heat of vaporization** [E M^{-1}] or MJ kg^{-1}
  - as **T_s increases, λ_v decreases**: $\lambda_v = 2.5 - 2.36 \times 10^{-3}\, T_s$
  - about **2.45 million joules** are required to evaporate 1 kg of water at 20°C

Sensible heat exchange (H_S)

- upward **sensible heat transfer**, H_S via turbulence:

$$H_S = K_h \cdot v_a (T_s - T_a)$$

$$K_h \equiv C_a \cdot \rho_a \left[6.25 \left(\ln \left[\frac{z_m - z_d}{z_0} \right] \right)^2 \right]^{-1}$$

  - K_h coefficient describing **upward transfer** of H_S by wind
  - C_a is **heat capacity** of vapour-bearing air

Bowen Ratio

- **Bowen Ratio** (β) used to describe ratio of H_S:LE

$$\beta \equiv \frac{H_S}{LE} = \frac{C_a \cdot \rho_a \cdot (T_s - T_a)}{0.622 \cdot \lambda_v \cdot (e_s - e_a)} = \gamma \cdot \frac{(T_s - T_a)}{(e_s - e_a)}$$

γ= **psychrometric constant**

- describes the heat capacity, air density and latent heat of vapourization properties of the air mass

Measuring & modeling ET

Five commonly used approaches:

1. **Direct measurement** of moisture loss
2. **Radiation balance-based**
3. **Aerodynamic based** (mass transfer)
4. **Combined radiation-aerodynamic**
5. **Temperature-based**

Direct measurement

- evaporation pan: E of exposed water from budget of W inputs & Δ storage volume (V)
  - $E_{pan} = P - [V_2 - V_1]$
  - more appropriate for short vegetation & ground cover
  - spatially limited, design biases, does not measure transpiration

Evaporation station at private laboratory of Robert Horton. In: Monthly Weather Review: 1919, Sept.: 608.

Direct measurement

- **Lysimeter**: Δ in weight of a control volume of soil proportionate to Δ in volume of moisture lost by surface evaporation & plant transpiration

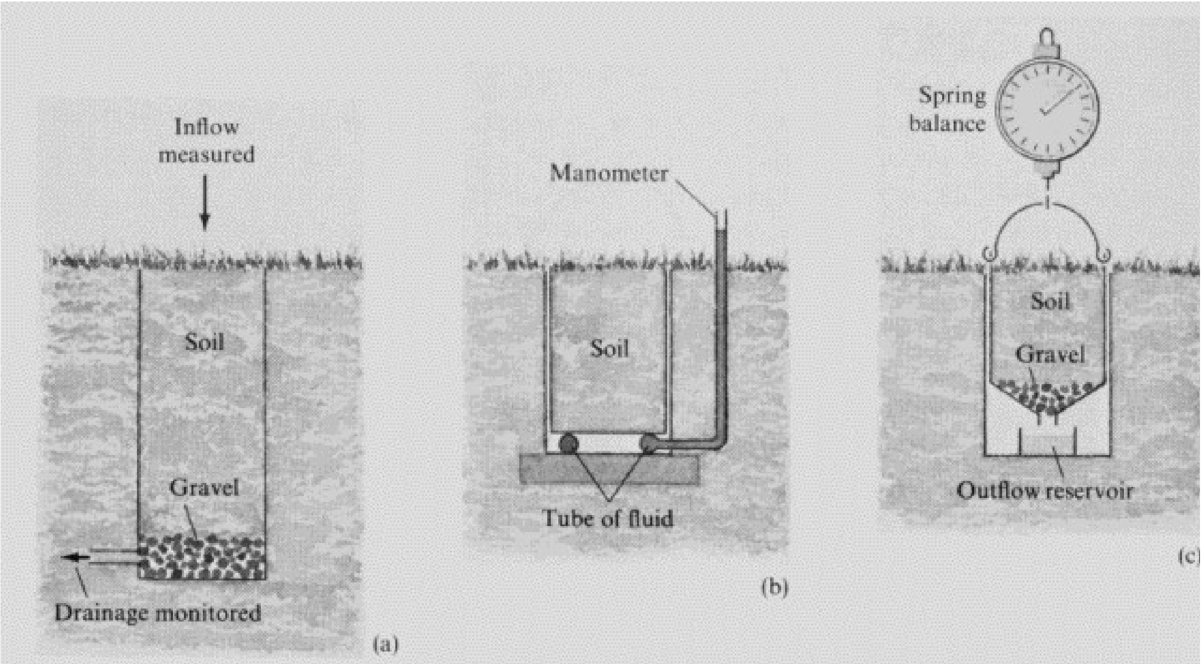
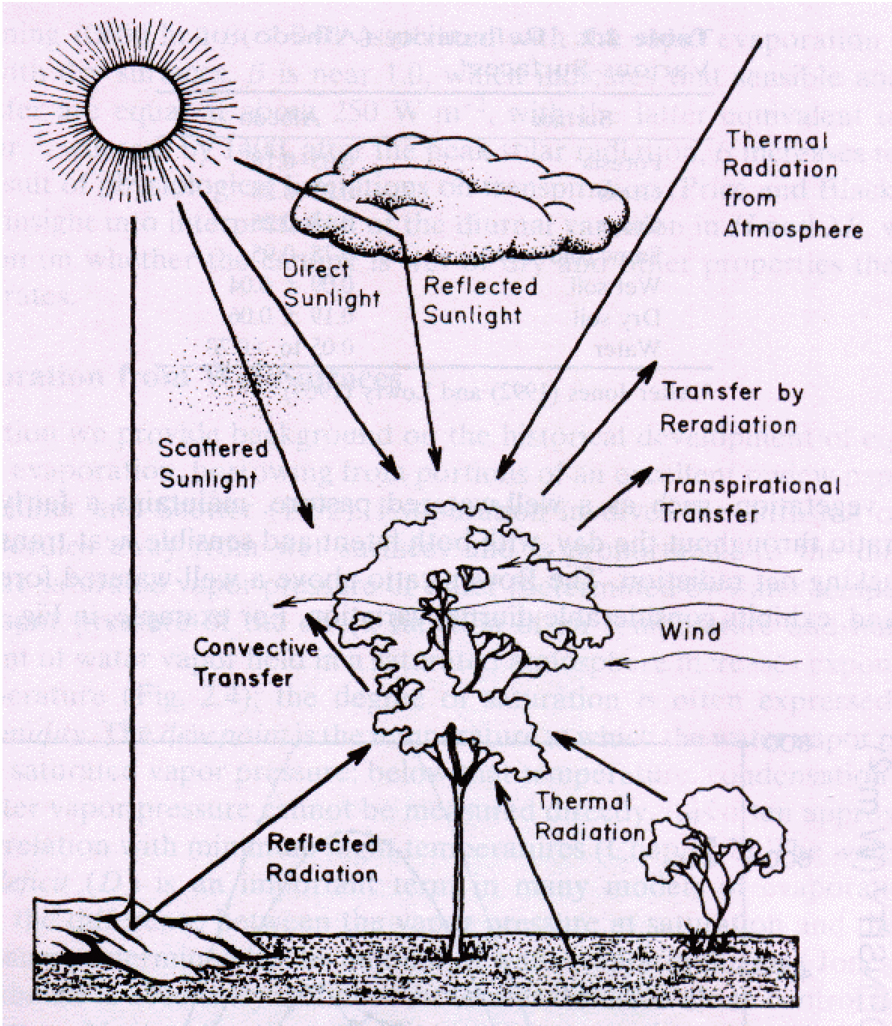


Figure 5-2 Schematic diagrams of types of lysimeters: (a) Drainage type. (b) Weighing float type. (c) Spring-balance weighing type.

(b) A giant 'floating' lysimeter containing a mature Douglas fir tree at Cedar River, Washington. The observer (centre left) is reading the manometer which monitors mass changes by the soil-tree monolith. Tensiometers are installed both inside and outside the lysimeter to ensure that similarity of moisture content is maintained. The brace in the foreground prevents rotation, and the tree is lightly 'guyed' to surrounding trees to prevent it falling over in high winds.

Radiation balance-based



Gates, D.M. 1980. Biophysical Ecology. Springer-Verlag, Berlin and New York.

Radiation balance-based

- We can describe the **net radiation** received by the Earth using the **Radiation Balance Equation**:

$$R_n = S_0(1.0-\alpha) + L_n$$

Where: S_0: Shortwave radiation from the Sun

α: Albedo (describing reflected rad'n)

L_n: Net longwave radiation

If $R_n > 0$, **net gain** of energy (daytime, summer)

$R_n < 0$, **net loss** of energy (nighttime, winter)

$R_n = 0$, then we have a **steady-state** condition

Radiation balance-based

- After the Earth's surface receives R_n radiative energy, the energy is **used in the following ways**:
- A portion of it will be used to **evaporate or transpirate** water from the liquid state to the gaseous state. This is called **latent heat (LE)** as the energy will be released when the gaseous water changes back to liquid state
- A portion of it will be used to **heat the atmosphere**, which is called **sensible heat (H_S)**
- A portion of it will **pass through the Earth's surface** to **heat the soil** below (Q)
- A **small fraction** of the energy is used by leaves for **photosynthesis** and this **energy is stored** in the chemical bonds of carbohydrate produced by photosynthesis (A)

Radiation balance-based

- We can describe the way the **net radiation** received by the Earth's surface is partitioned using the **Energy Balance Equation**:

$$R_n = LE + H_S + H_G + A$$

Where:		
	LE:	Latent heat
	H_S:	Sensible heat
	H_G :	Energy stored in the soil
	A:	Energy stored in photosynthate

- How R_n is distributed among the items on the right hand side is determined by the **ecosystem biophysical characteristics** and has major consequences for ecosystem development and functions

Radiation balance-based

- $LE = R_S + R_L - H_G - H_S + H_A - (\Delta H/\Delta t)$
  - R_S = net **short-wave** radiation
  - R_L = net **long-wave** radiation
  - H_G = net **ground heat** conduction (typically small)
  - H_S = net **sensible heat** output to atmosphere
  - H_A = net input associated with **inflow/outflow of water** (advected energy)
  - ΔH = Δ heat **stored in evaporating body** (per unit area)

  - additional **photosynthetic energy** term (A) can be added... may amount to up to <u>3%</u> of R_N

Radiation balance-based

- **Energy balance** of a surface
  - $R_N = H_G + H_S + LE$

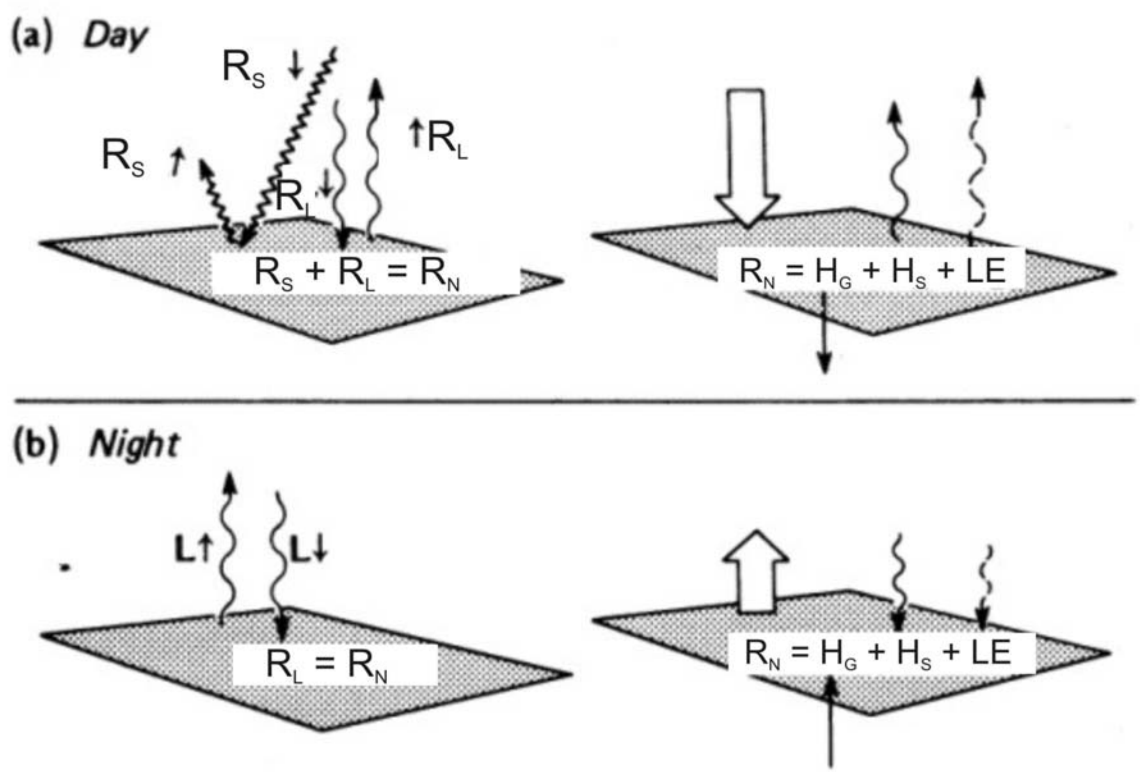


Figure 1.11 Schematic summary of the fluxes involved in the radiation budget and energy balance of an 'ideal' site, (a) by day and (b) at night.

Radiation balance-based

Radiation balance-based

- You can calculate the ratio between sensible and latent heat fluxes, and this is known as the **Bowen Ratio (β)**:

$$\beta = H / LE$$

- The sensible heat flux is often **difficult to measure**, but if you can estimate the Bowen Ratio, you can rewrite the **net radiation balance equation** in terms of latent heat:

$$R_n = H + LE + H_G$$

$$R_n = (\beta * LE) + LE + H_G$$

$$LE = (R_n - H_G) / (1 + \beta)$$

Radiation balance-based

- Evaporation calculated via **Bowen ratio energy balance** method

$$LE = R_N - H_S - H_G$$

- Method seeks to apportion available energy between sensible and latent heat flux by considering their ratio

$$\beta = \frac{H_S}{LE} = \gamma \cdot \frac{\Delta T}{\Delta e}$$

$$LE = \frac{(R_N - H_G)}{(1+\beta)}$$

- assumes neutral stability (buoyancy effects are absent) and steady state (no marked shifts in radiation)

$$E = \frac{(R_N - H_G)}{\rho_W \cdot \lambda_V \cdot (1+\beta)}$$

Radiation balance-based

Subsurface heat flux for a lake

$$H_G = H_{GL} + H_{GB}$$

H_G = total subsurface heat flux (W m^{-2})
H_{GL} = lake heat storage (W m^{-2})
H_{GB} = heat conduction into lake bed (W m^{-2})

$$H_{GL} = C_w \times \frac{\Delta T}{\Delta t} \times \frac{V}{A}$$

C_w = heat capacity of water (J m^{-3} kg^{-1})

ΔT = change in temperature (°C) over time step Δt (sec)

V= lake volume (m^3) & A is lake area (m^2)

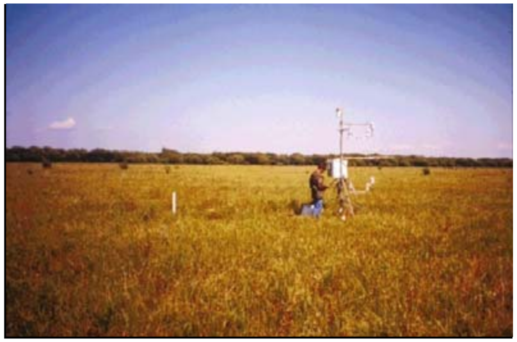
$$H_{GB} = -K_B \times \frac{(T_1 - T_2)}{\Delta z}$$

K_B = bed thermal conductivity (W m^{-1} °C^{-1}) estimated from bed grain size

T_1 is temperature (°C) at water bed interface and T_2 at some depth below lake

Δz = distance (m) between T_1 and T_2

Radiation balance-based



Subsurface heat flux for **soil**

$$H_G = C \times \frac{\Delta T}{\Delta t} \times \Delta z$$

$$C = x_m C_m + x_{om} C_{om} + x_w C_w + x_a C_a$$

Where:

T = soil temperature

t = time

z = depth of soil for which measurements are taken

C = volumetric heat capacity

x = fraction of soil constituent (mineral (m), organic matter (om), water (W) and air (a))

Energy fluxes change over time

Wetland adjacent to Lake Ontario

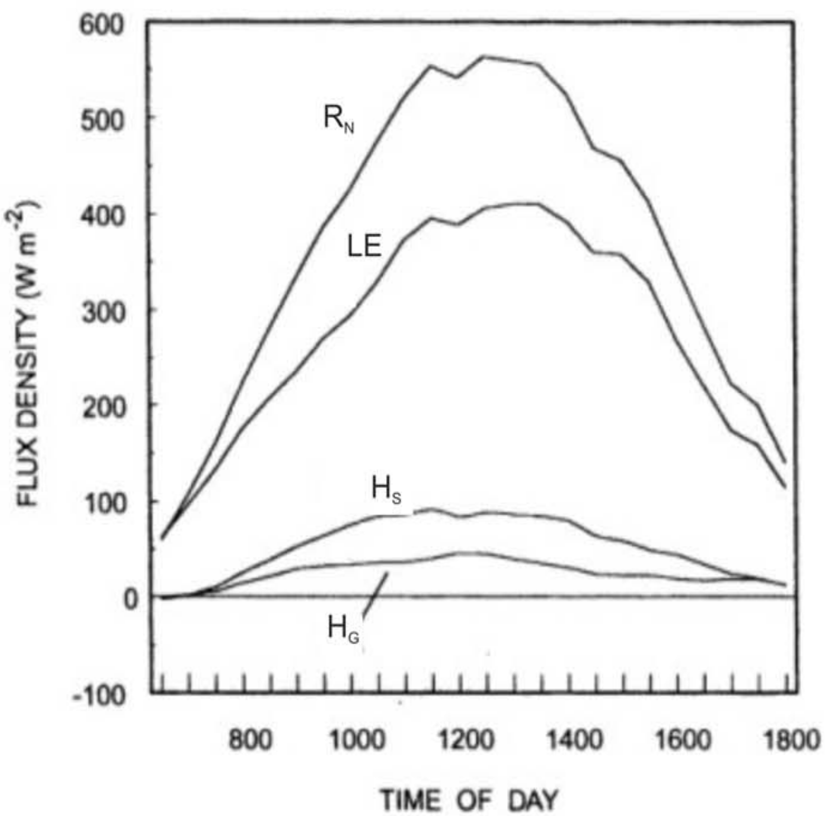


Fig. 2. Averages of the half-hourly measurements of net radiation (R_N), and latent (LE), sensible (H_S) and ground (H_G) heat flux. The energy balance components were measured over the *Typha* canopy between 14 June and 6 August 1991.

Energy fluxes change over time

Experimental Field in California

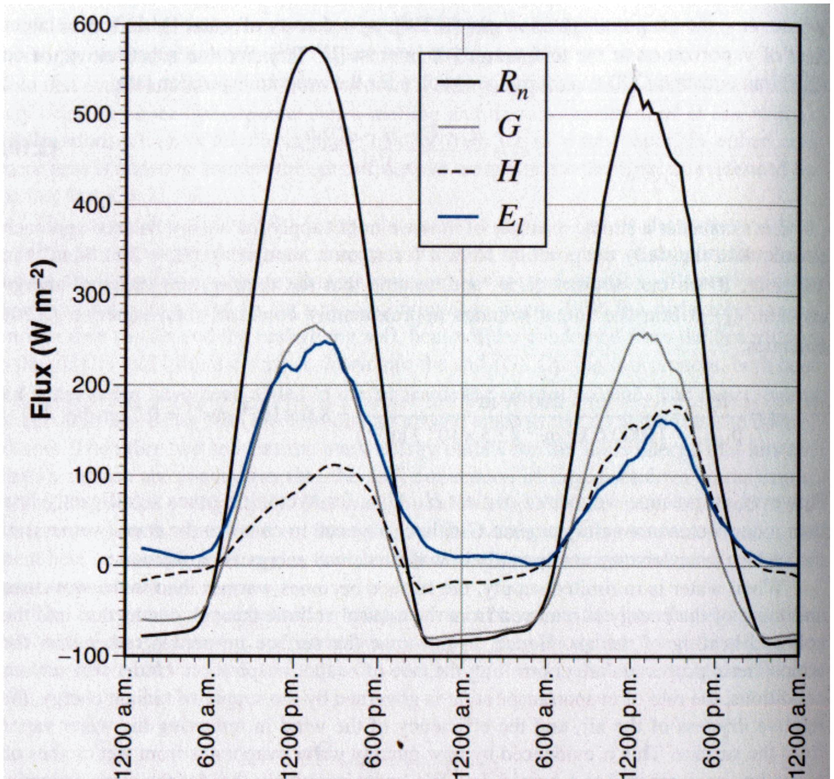


Figure 2.10 Measured energy fluxes from an experimental field in California. As the surface dries during the experiment, the latent heat flux (E_l) is reduced and the sensible heat flux (H) increased. Data courtesy of John D. Albertson.

Hornberger et al. 1998. Elements of Physical Hydrology. The Johns Hopkins University Press, Baltimore and London.

Energy fluxes change over time

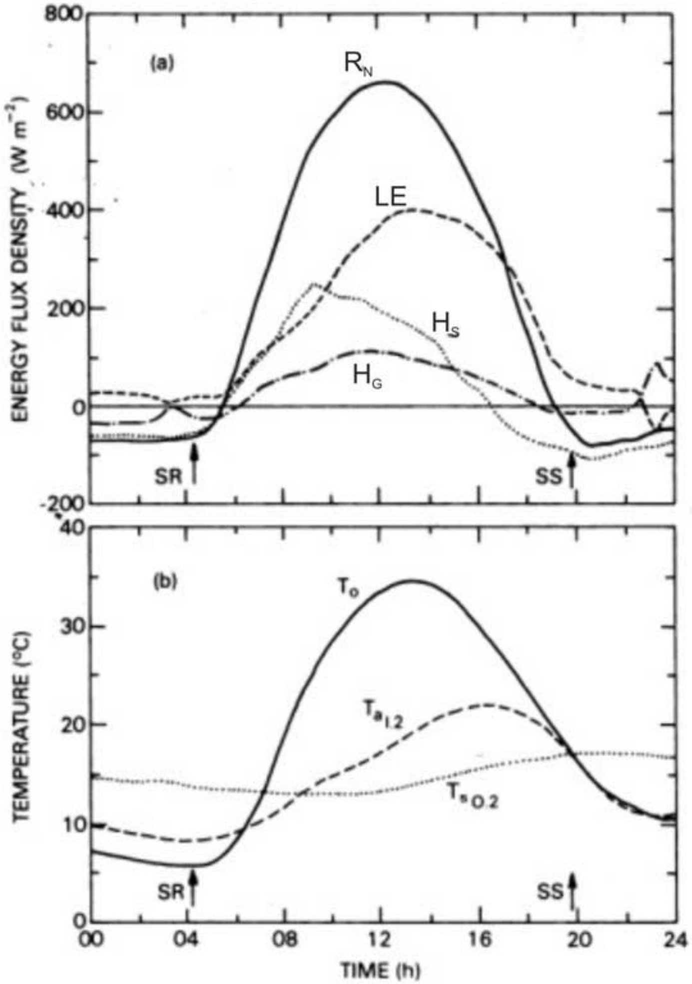


Figure 1.10 (a) Energy balance components for 30 May 1978 with cloudless skies at Agassiz, B.C. (49°N) for a moist, bare soil, and (b) temperatures at the surface, in the air at a height of 1·2 m and in the soil at a depth of 0·2 m (after Novak and Black, 1985). The following table gives the energy totals for the day (MJ m^{-2} day^{-1}).

Energy balance for a moist, bare soil on May 30, 1978 at Agassiz, BC.

Daily summary

$R_N = 18.0$

$H_S = 2.3$

$LE = 13.4$

$H_G = 2.3$

units in MJ m^{-2} day^{-1}

note: 1 W = 1 J s^{-1}

Radiation balance-based

- Semi-empirical approach requires measurements on one level above surface

Priestly & Taylor (1972):

$$\mathrm{PET} = \alpha \cdot \frac{s}{s+\gamma} \cdot \frac{(R_N - \mathrm{H_G})}{\rho_w \cdot \lambda_v}$$

where:

PET is **potential evapotranspiration** (mm per time)

$s = (e^*_s - e^*_a)/(T_s - T_a)$... describes **gradient** of e* vs. T at a given air temperature

α is an empirically derived **evaporability factor** (usually 1.26)

Aerodynamic Profile Method

Unmodified method applies to the following **restricted conditions**:

i) **neutral stability** - buoyancy effects absent
ii) **steady state** - no marked shifts in radiation or wind fields during observation periods
iii) constancy of fluxes with height - **no vertical divergence or convergence**
iv) **similarity** of all transfer coefficients

under these conditions the logarithmic wind profile is valid and the wind gradient is found to be **inversely proportional to the height above a surface**

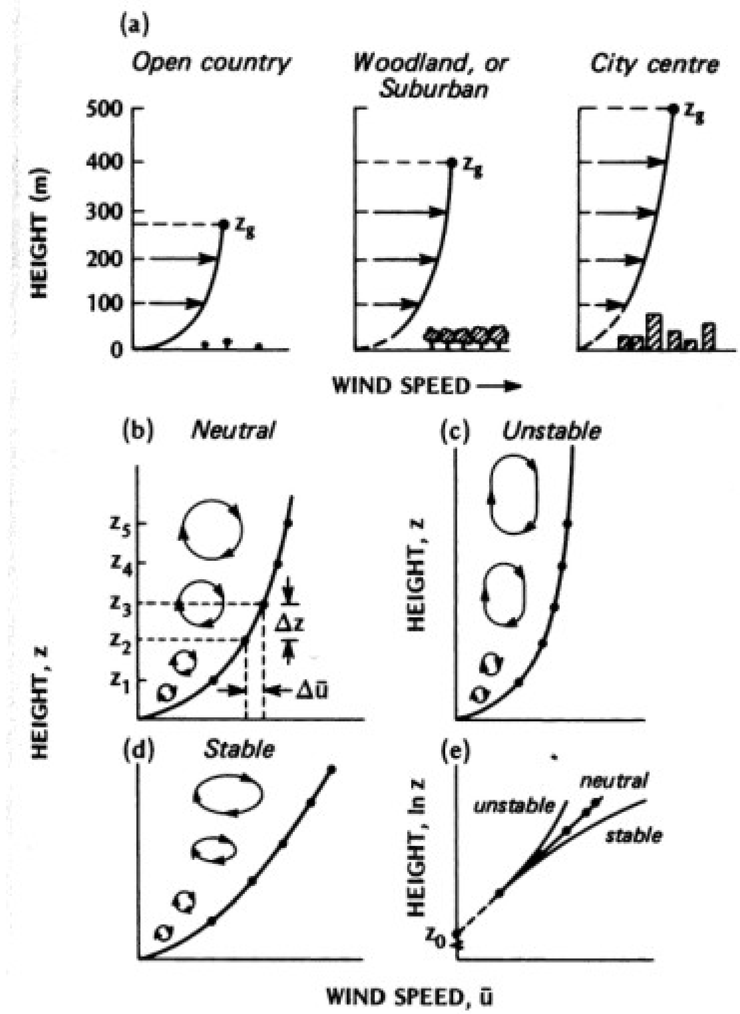


Figure 2.10 The wind speed profile near the ground including: (a) the effect of terrain roughness (after Davenport, 1965), and (b) to (e) the effect of stability on the profile shape and eddy structure (after Thom, 1975). In (e) the profiles of (b) to (d) are re-plotted with a natural logarithm height scale.

Aerodynamic Profile Method

$$E = -\ \rho_a \cdot \kappa^2 \cdot \frac{(u_2 - u_1)\cdot(q_2 - q_1)}{(\ln z_2/z_1)^2} \cdot (\phi_m \times \phi_v)^{-1}$$

where:

κ is the von Karman constant (0.4)

u is the mean windspeed (m s^{-1}) at height z

q is the mean specific humidity (g kg^{-1})

z is the measurement height above surface

1= lower and 2 = upper

φ_m and φ_v are stability corrections for momentum and water vapor

Combined approaches

- **Penman** (1948) developed a method considering the factors of **both energy supply and turbulent transport of water vapor** from an evaporating surface

- Requires meteorological measurements at **only 1 level**

- in the combination method LE is **calculated as the residual** in the energy balance equation with sensible heat flux estimated by means of **an aerodynamic equation**

- **widely used** for estimating potential evapotranspiration

- original method designed to estimate evaporation from **open-water or well-watered surfaces**
  - e.g., lake, pond, and wetlands

Types of ET

TABLE 7-1
Classification of Types of Evapotranspiration

Evapotranspiration Type	Type of Surface	Availability of Water to Surface	Stored Energy Use	Water-Advected Energy Use
Free-water evaporation[a]	Open water	Unlimited	None	None
Lake evaporation	Open water	Unlimited	May be involved	May be involved
Bare-soil evaporation	Bare soil	Limited to unlimited	Negligible	None
Transpiration	Leaf or leaf canopy	Limited	Negligible	None
Interception loss	Leaf or leaf canopy	Unlimited	Negligible	None
Potential evapotranspiration	Reference crop[b]	Limited to air, unlimited to plants	None	None
Actual evapotranspiration	Land area[c]	Varies in space and time	Negligible	None

[a] Also called **potential evaporation.**

[b] Usually a complete ground cover of uniform short vegetation (e.g., grass); discussed further in Section 7.7.1.

[c] May include surface-water bodies and areas of bare soil.

Combined approaches

Penman (1948):

$$\mathrm{PET} = \frac{s \cdot (\mathrm{R_N} - \mathrm{H_G}) + \dfrac{\mathrm{C_a} \cdot (\mathrm{e_s} - \mathrm{e_a})}{\mathrm{r_a}}}{\rho_\mathrm{W} \cdot \lambda_\mathrm{V} \cdot (s + \gamma)}$$

$$\mathrm{r_a} = \frac{[\ln(\mathrm{z} - \mathrm{d})/\mathrm{z_O}]^2}{\kappa^2 \cdot \mathrm{u_z}}$$

Aerodynamic resistance, r_a, describes the resistance from the water or vegetation upward and involves friction of air flowing over water or vegetative surface

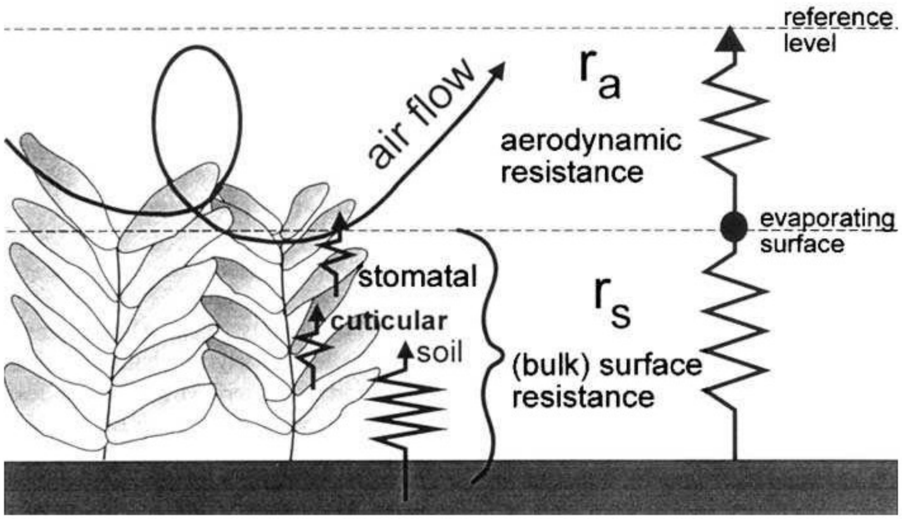
r_a = aerodynamic resistance (s m^{-1})

u_z = is wind speed (m s^{-1}) at elevation z (m)

κ= van Karman's constant (0.4)

κz_o = roughness length (m)

κd = zero plane of displacement



Actual evapotranspiration (ET)

- where PET > available moisture, ET is **water-limited**
  - e.g., hot arid regions
  - ET also limited by **insufficient energy** to fuel the process (i.e., ET = PET)… e.g., arctic environments
- how to estimate ET?
  - **water budget approach** $ET = W / ([1+(W/PET)^2]^{1/2})$
  - **soil-moisture function** $ET = F(\theta_{rel}) \bullet PET$
    - $F(\theta_{rel})$ is water infiltrated to relative water content (θ_{rel} in mm) where $\theta_{rel} = (\theta - \theta_{pwp} / \theta_{fc} - \theta_{pwp})$ where 'pwp' = permanent wilting point & 'fc' = field capacity
  - **weighing lysimeter** Δ in weight of a control volume of soil proportionate to Δ in volume of moisture lost by surface evaporation & plant transpiration

Combined approaches

- **Penman-Monteith equation** common for ET from a vegetated land surface

$$PET = \frac{s \cdot (\mathrm{R}_N - \mathrm{G}) + \dfrac{\rho_a c_a \cdot (e_s - e_\mathrm{a})}{r_a}}{\rho_w \cdot \lambda_v \cdot [s + \gamma \cdot (1 + r_c / r_a)]}$$

$$r_c = \frac{\mathrm{r_i}}{LAI_{active}}$$

where:

r_c = canopy resistance (s m^{-1})

r_i = bulk stomatal resistance of the well-illuminated leaf (s m^{-1})

LAI_{active} = active (sunlit) leaf area index (m^2 leaf area per m^2 soil surface)

Temperature-based

- **Thornthwaite** (1948) related PET to monthly temp & daylength

$$PET = 16 \times \left(\frac{\mathrm{h}}{12}\right) \times \left(\frac{\mathrm{N}}{30}\right) \times \left(\frac{10 \times T_{ma}}{I}\right)^{a}$$

$$I = \sum_{\mathrm{i}=1}^{12} \left(\frac{T_{ma}}{5}\right)^{1.5}$$

$$a = 0.49 + 0.0179 - 0.000771 \times I^{2} + 0.000000675 \times I^{3}$$

where:

PET = potential evapotranspiration (mm mo^{-1})

h = day length hours

N = number of days in the month

T_{ma} = mean monthly air temperature (°C)

I = annual heat index

Temperature-based

Hamon (1963) see Dingman (2002)

$$PET = 29.8\,D\left[\frac{e_a^*(T_a)}{T_a + 273.2}\right]$$

where:

PET = potential evapotranspiration (mm d^{-1})

D = hours of sunshine in 12 hour units (hr)

T_a = air temperature (^oC)

$$e_a^*(T_a) = 0.611\exp\left(\frac{17.27\,T_a}{T_a + 237.3}\right)$$

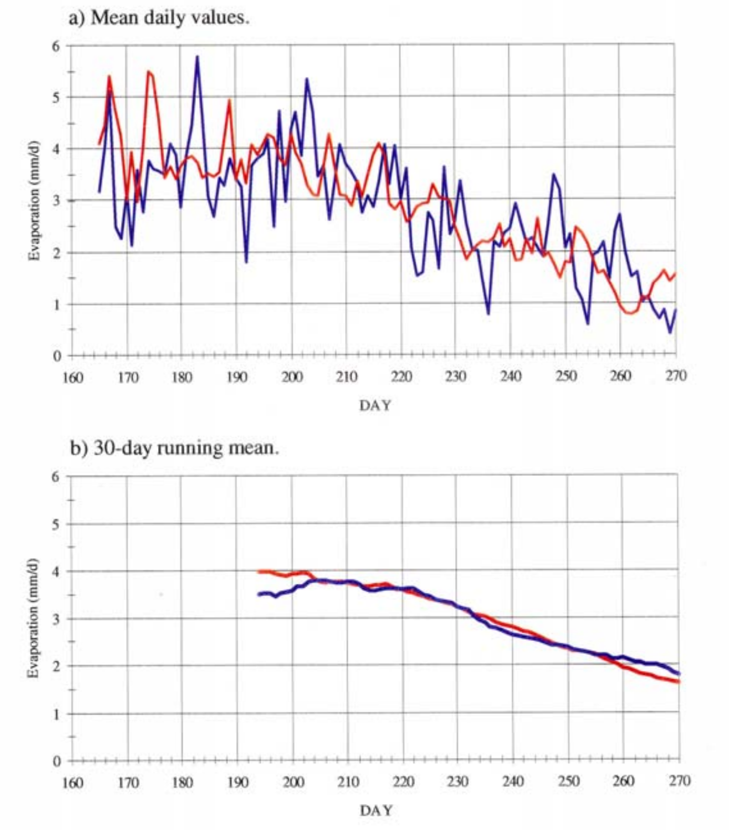
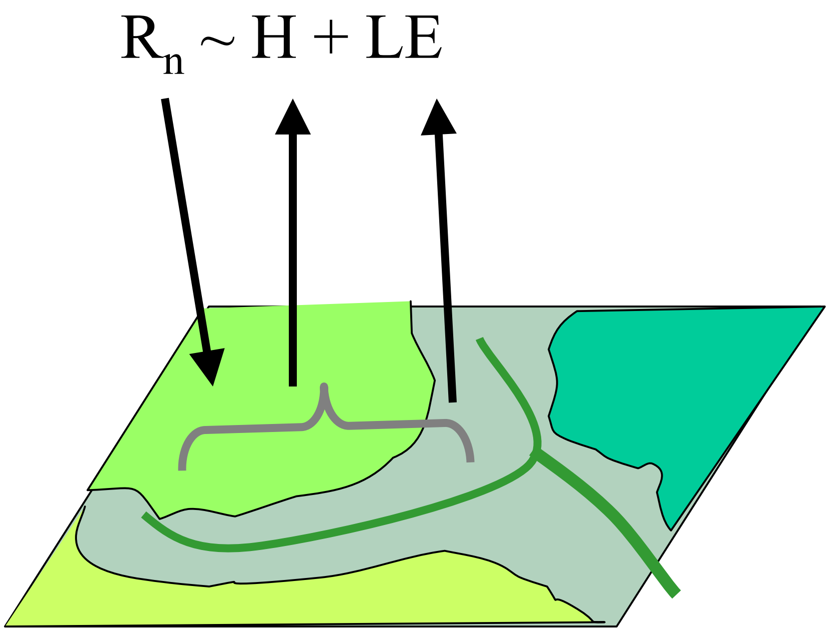


Figure 13. Comparison of the Hamon (red line) to the Priestley-Taylor (blue line) method of estimating open-water evaporaion.

Surface water/energy budget coupling over heterogeneous terrain



$$LE = f_{veg}\, LE_{veg} + (1 - f_{veg})\, LE_{soil}$$

$$LE = f(R_n, T, g_c, g_a, g_{soil}, VPD)$$

$$g_a = f(\text{canopy structure, wind, ...})$$

$$g_c = f(\text{soil water, VPD, PAR, T, LAI})$$

$$g_{soil} = f(\text{soil water, ...})$$

T_s lower with greater LE (evaporative cooling) as a function of soil water (other factors), greater canopy cover (higher NDVI)

T_s and NDVI estimated by a set of operational remote sensors