Evapotranspiration
What is evapotranspiration?

Evapotranspiration (ET) is a collective term for *all* processes through which water in liquid or solid form becomes atmospheric water vapor. It includes evaporation from bare soil, lakes and rivers and vegetative surfaces. It also includes transpiration, which represents evaporation from within the leaves of plants through stomatal openings.


http://evolution.berkeley.edu/evolibrary/article/mcelwain_02
The importance of ET

Globally, 62% of precipitation that falls over continents is evapotranspirated (72,000 km³ yr⁻¹, or 37 cm per unit area over continents)

• 97% over land (evaporation and transpiration)
• 3% over open water on continents (evaporation)

ET exceeds runoff \( R \) on all continents except Antarctica.

ET is the primary link in the global hydrologic cycle between the land and the atmosphere. It plays a key role in runoff and water availability and agriculture. Most of the world’s food supply is grown on irrigated land and efficient irrigation requires knowledge of transpiration.
Requirements for evapotranspiration

To enable and maintain evapotranspiration, you need:

(1) Available liquid water;
(2) Energy to break hydrogen bonds;
(3) A vertical vapor pressure gradient;
(4) Turbulence to whisk away vapor molecules to maintain the vertical vapor pressure gradient.

Physics of evaporation

Consider a body of water, such as a lake. Air with a temperature $T_a$ lies above the water surface with temperature $T_w$. Molecules at the water surface are attracted to each other by hydrogen bonds, but some have sufficient energy to sever the bonds and enter the air; the number of molecules with sufficient energy to break the hydrogen bonds is proportional to $T_w$. To increase $T_w$ you have to *add energy*. Molecules accumulate in the layer immediately above the surface and an equilibrium develops, with as many molecules re-entering the water as there are escaping. In the case of a water body, this thin layer is saturated, with a vapor pressure $e_s$ at temperature $T_s$. The rate of evaporation is the rate at which molecules enter the unsaturated layer above with a lower vapor pressure $e_a$ at temperature $T_a$.

Dalton’s Law:

$$E \propto e_s - e_a$$

Hence, the drier the airmass above the surface, the bigger the vertical vapor pressure gradient, and the greater the evaporation, other factors being equal.

Dingman 2002, Figure D-4
Transpiration involves absorption of soil water by roots, movement of liquid water through the plant’s vascular system of the roots, stem and branches to the leaves, and movement through the vascular system of the leaves to the walls of stomatal cavities where evaporation takes place. Water vapor in these cavities then moves into the ambient air through leaf openings called stomata. Plants live by absorbing CO₂ to make carbohydrates; the stomatal cavities provide a place where CO₂ can be dissolved in water to enable entry into the plant tissues. Air in stomatal cavities is saturated at the leaf temperature, and water moves from the cavities to the ambient air due to vapor pressure differences. Plants can exert some control over the size of stomatal openings and hence evaporation.
Saturation vapor pressure, relative humidity and dewpoint

The saturation vapor pressure $e^*$ depends on temperature $T$. It can be closely approximated as:

$$e^* = 0.611 \cdot \exp\left(\frac{17.3 \cdot T}{T+237.3}\right)$$

With temperature in $^\circ$C. The relative humidity $RH$ is defined as the ratio of the actual vapor pressure of the air at temperature $T$ to the saturation value at temperature $T$:

$$RH = \frac{e}{e^*}$$

The dewpoint $T_d$ is the temperature to which an air parcel must be cooled to get to saturation; further cooling results in condensation (however, under certain conditions, the air can be supersaturated, as discussed earlier, this is important for precipitation).

http://qwickstep.com/search/saturated-vapor-pressure.html
Physics of evaporation (cont.)

• Evaporation *does not require* that the surface layer be at saturation (as would be the case over a free water or ice surface or within a stomata cavity); what matters is the *vertical gradient of vapor pressure*.

• Depending on $T_s$, $T_a$ and $e_s$ and $e_a$, the difference between the two vapor pressures ($e_s - e_a$) can be positive (evaporation is occurring), negative (deposition is occurring) or zero (neither evaporation or condensation are occurring in a net sense)

• The value of $e_a$ can be less than or equal to the saturation vapor pressure at $T_a$; if equal, the relative humidity is 100%. Saturation at $T_a$ does not mean that no evaporation can occur; what counts is the vertical gradient in vapor pressure.

• Evaporation occurs in the same way over an ice surface as over a water surface. However, the saturation vapor pressure over and ice surface is slightly lower than over a water surface
Physics of evaporation (cont.)

The latent heat of vaporization $\lambda_v$ (2.501x10$^6$ J kg$^{-1}$) represents the energy that is required to break the hydrogen bonds. A unit mass of vapor at the same temperature as a unit mass of liquid water has a higher energy content because it took energy to break the hydrogen bonds and the vapor mass possesses this extra energy. If vapor returns to the water phase, this energy is released.

The latent heat flux $LE$ (W m$^{-2}$ = J s$^{-1}$ m$^{-2}$) and $E$ (kg m$^{-2}$ s$^{-1}$) are hence directly coupled as the mass transfer is accompanied by an energy transfer, specifically

$$LE = \lambda_v . E$$

If the surface is snow or ice, the latent heat transfer requires an additional energy component - the latent heat of fusion $\lambda_f$ (3.337 x 10$^5$ J kg$^{-1}$)

$$LE = (\lambda_v + \lambda_f) . E$$

Note that 1 kg m$^{-2}$ of evaporation is the same as a 1 mm of water over a square meter (assuming a water density of 1000 kg m$^{-3}$).
As previously stated, the drier the airmass the greater the potential vertical vapor gradient and the greater the evaporation, other factors being equal. However, given that there is a vapor transfer from higher to lower concentration (the transfer is “down the gradient”), the gradient would get smaller and smaller with time, and eventually go away, essentially gumming up the works. Hence, to maintain evaporation, we need a processes to whisk evaporated water molecules away; so as to maintain a vertical gradient. This is why $E$ and the latent heat flux $LE$ depend so critically on the turbulent wind field. Stated most simply:

$$E = K_E v_a (e_s - e_a)$$

Where $v_a$ is the horizontal wind speed and the coefficient $K_E$ reflects the efficiency of vertical moisture transport by turbulent eddies. Approaches to estimating $LE$ (and hence $E$) that rely on knowledge of turbulence were visited earlier under discussion of estimating snowmelt. See Chapter 7 in Dingman (2002) for further details.
Bowen Ratio

The Bowen ratio (B), the ratio of the sensible heat flux to the latent heat flux, can be calculated without actually measuring the fluxes:

\[ B = \frac{H}{LE} \]

\[ B = \frac{c_a \cdot \rho_a \cdot (T_s - T_a)}{[0.622 \cdot \lambda \cdot (e_s - e_a)]} \]

Where \( c_a \) is the specific heat capacity of air (1.00x10^{-3} MJ kg^{-1} K^{-1}) and \( \rho_a \) is the air density. The Bowen ratio hence depends on the ratio of the vertical difference in air temperature and vapor pressure.

The term \( \gamma = \frac{c_a \cdot P}{0.622 \cdot \lambda \cdot v} \)

is the the psychrometric content, where \( P \) is atmospheric pressure. The constant 0.622 is the ratio of the molecular weights of air and water vapor. Hence:

\[ B = \gamma \cdot \frac{(T_s - T_a)}{(e_s - e_a)} \]

The evaporative fraction is

\[ EF = \frac{LE}{(LE + H)} = \frac{1}{(B + 1)} \]
Physics of evaporation (cont)

As discussed, evaporation requires energy to break the hydrogen bonds. The general energy balance for an evapotranspiring body (a lake, a given volume of soil with or without vegetation) is:

\[
LE = K + L - G - H + A_w - \Delta Q/\Delta T
\]

\[
E = (K + L - G - H + A_w - \Delta Q/\Delta T)/\rho_w \cdot \lambda_v
\]

LE has units of \( J \, m^{-2} \) (the energy flux per unit surface area of the evapotranspiring body) and \( E \) has units of \( m \). \( K \) is net solar radiation, \( L \) is the net longwave radiation, \( G \) is conduction into the ground, \( H \) is the turbulent sensible heat flux, \( A_w \) is the heat input from influx and outflows of water (water advected energy) and \( \Delta Q/\Delta T \) is the time change in the heat stored in the body per unit area between the beginning and the end of \( \Delta T \). Fluxes are positive when they represent an energy source for \( LE \).

**Note 1**: This is just an energy budget. Be careful in over-interpreting the budget as “energy available for \( LE \) or \( E \)”, for evaporation itself has a cooling effect on the surface temperature, which then affects other terms such as \( H \) and \( L \).

**Note 2**: The importance of each energy term depends on the type of evapotranspiring body. For example, for a transpiring leaf, the key energy input is solar radiation (supporting photosynthesis). \( G \) is absent and having little mass, the use of stored energy is minimal. By contrast, stored energy use can be very important for lake evaporation.
Physics of evaporation (cont).

$$LE = K + L - G - H + A_w - \frac{\Delta Q}{\Delta T}$$

$K$ the net solar radiation flux at the surface ($K_{in} - K_{out}$), is either zero or positive (downwards); positive $K$ is an energy source supporting $LE$. 

$L$, the net longwave radiation flux at the surface ($L_{in} - L_{out}$), is usually negative (upwards), which hence has a dampening effect on $LE$. However, $LE$ also affects $L$ and $H$!

$H$, the turbulent sensible heat flux, is an energy source for $LE$ when directed downwards (towards the surface, negative $H$ in the above equation). This would be the case with a temperature inversion. More typically, $H$ is directed upwards, away from the surface.

$G$ is conduction into or out of the ground; if the conduction is into the volume (negative $G$ in the above equation), this is an energy source for $LE$. 

$A_w$
Consider a few different situations

**Boulder Colorado after a July thunderstorm:** Clouds clear and there is a strong input of solar radiation to the surface. There is plenty of water available to evaporate, but the airmass itself is quite dry. It is windy. Evapotranspiration is big.

**Coastal Maine in July:** There is plenty of water at the surface to evaporate. However, it is a much cloudier place than Colorado, so there is less solar radiation at the surface. While it is also often a windy place, the prevailing airmass is much moister. Evapotranspiration is hence smaller than in Colorado in July after a thunderstorm.

**The Norwegian Sea (Arctic) in winter:** The airmass is cold and dry. It is a very windy. While there is very little solar radiation, the ocean surface is ice-free, and quite warm with respect to the latitude of about 75°N. This is because of the poleward advection of ocean heat by currents (lots of water advected energy). Evaporation (no transpiration!) is big.
Classification of types of evapotranspiration

<table>
<thead>
<tr>
<th>Evapotranspiration Type</th>
<th>Type of Surface</th>
<th>Availability of Water to Surface</th>
<th>Stored Energy Use</th>
<th>Water-Advected Energy Use</th>
</tr>
</thead>
<tbody>
<tr>
<td>Free-water evaporation</td>
<td>Open water</td>
<td>Unlimited</td>
<td>None</td>
<td>None</td>
</tr>
<tr>
<td>Lake evaporation</td>
<td>Open water</td>
<td>Unlimited</td>
<td>May be involved</td>
<td>May be involved</td>
</tr>
<tr>
<td>Bare-soil evaporation</td>
<td>Bare soil</td>
<td>Limited to unlimited</td>
<td>Negligible</td>
<td>None</td>
</tr>
<tr>
<td>Transpiration</td>
<td>Leaf or leaf canopy</td>
<td>Limited</td>
<td>Negligible</td>
<td>None</td>
</tr>
<tr>
<td>Interception loss</td>
<td>Leaf or leaf canopy</td>
<td>Unlimited</td>
<td>Negligible</td>
<td>None</td>
</tr>
<tr>
<td>Potential evapotranspiration</td>
<td>Reference crop&lt;sup&gt;b&lt;/sup&gt;</td>
<td>Limited to air, unlimited to plants</td>
<td>None</td>
<td>None</td>
</tr>
<tr>
<td>Actual evapotranspiration</td>
<td>Land area&lt;sup&gt;c&lt;/sup&gt;</td>
<td>Varies in space and time</td>
<td>Negligible</td>
<td>None</td>
</tr>
</tbody>
</table>

<sup>a</sup>Also called potential evaporation.

<sup>b</sup>Usually a complete ground cover of uniform short vegetation (e.g., grass); discussed further in Section 7.7.1.

<sup>c</sup>May include surface-water bodies and areas of bare soil.

Dingman 2002, Table 7-1

One can define various evapotranspiration “types” that depends on:

- Type of surface (open water, such as a lake, bare soil, leaf or leaf canopy, a reference crop)
- Availability of water (unlimited, such as over a lake, limited, or variable)
- Stored water energy use (is energy in a given volume used to promote evaporation)
- Water advected energy use

**Note 1:** Free water evaporation is evaporation that would occur from an open water surface in the absence of advection and change in heat storage

**Note 2:** The type of evapotranspiration bears on the method of estimation.
Free water, lake and wetland evaporation

The availability of water at the surface is unlimited. Water advected energy and use of stored heat can be important.

Estimation strategies include:

1) Water balance approach (compute as residual)
2) Mass transfer (profile methods, eddy correlation, we’ve already talked about these at some length)
3) Energy budget (already introduced, we’ll build)
4) Penman approach
5) Pan evaporation
Water balance approach

\[ E = (P + R_{in} + G_{in}) - (G_{out} + R_{out}) \]

The challenge: estimate E from a lake as a residual from other measured terms. To do so, you need both the runoff R into and out of the lake, potentially requiring many stream gauges. As with a watershed, it may be hard to get a good estimate of the average precipitation. If you are interested in storage changes, you need good information on lake bathymetry.

http://www.paddletrips.net/moosehead_lake.htm
Limitations of the water balance approach

The figure at right shows the water balance for the Williams Fork Reservoir, Colorado, for four years. The key point is that the uncertainty in the overall budget (vertical bars, standard deviation of the total budget) is large compared to the annual evaporation. Using the water balance approach as a check on evaporation estimated from other means is obviously problematic.

Dingman 2002, Figure 7-1
Recall the generic equation for the energy budget:

\[
LE = K + L - G - H + A_w - \Delta Q/\Delta T
\]

\[
E = (K + L - G - H + A_w - \Delta Q/\Delta T)/\rho_w \cdot \lambda_v
\]

Measuring all of the terms presents a formidable challenge. It may be possible to eliminate some small terms. For example, for a lake, heat conduction into/from the ground (G) can be neglected. The turbulent sensible heat flux is hard to measure (one needs wind data), but can be eliminated using the Bowen ratio B:

\[
B = \gamma \cdot (T_s - T_a)/(e_s - e_a)
\]

\[
H = B \cdot LE = B \cdot \rho_w \cdot \lambda_v \cdot E \quad \text{and by substitution}
\]

\[
E = (K + L + A_w - \Delta Q/\Delta T)/[(\rho_w \cdot \lambda_v \cdot (1+B)]
\]

However, one still needs data for surface and air temperature and humidity.
Many empirical relationships have been developed to get the radiation terms. The incoming solar radiation flux $K_{in}$ can be estimated from the clear sky incoming flux $K_{cs}$ (which can be calculated fairly accurately) and the fraction of the sky covered with cloud (C). For example:

$$K_{in} \approx [0.355 + 0.68 \cdot (1-C)] \cdot K_{cs}$$

To get the net shortwave flux $K$, one then assumes a surface albedo.

The net longwave flux can be estimated from Stefan-Boltzmann relationships using the air temperature $T_a$, empirical relationships for the longwave emissivity of the atmosphere (depending on vapor pressure and cloud cover); and an assumed surface emissivity. For example, with temperatures in °C (see equations 5-34 to 5-37 in Dingman 2002), we have

$$L \approx \epsilon_w \cdot \epsilon_{at} \cdot \sigma \cdot (T_a + 273.2)^4 - \epsilon_w \cdot \sigma \cdot (T_s + 273.4)^4$$

One can also use empirical approaches to estimate humidity.
Water advected energy

For a lake, water advected energy can be important. Consider the heat content of all of the water flows into and out of the lake. If there is more going in than out, the water advected energy is positive, which means more energy for evaporation.

\[ A_w = c_w \rho_w (w \cdot T_a + SW_{in} \cdot T_{swin} - SW_{out} \cdot T_{swout} + GW_{in} \cdot T_{gwin} - GW_{out} \cdot T_{gwout}) \]

Where \( \rho_w \) is the mass density of water, \( c_w \) is the specific heat of water, \( w \) is the precipitation rate, \( SW \) and \( GW \) represent surface water and groundwater inflows (in) and outflows (out) expressed as volumes per unit area, and the \( Ts \) represent the respective temperatures.

Change in energy storage

The change in energy storage can also be important for a lake:

\[ \Delta Q = (c_w \rho_w / A_L) \cdot (V_2 \cdot T_{L2} - V_1 \cdot T_{L1}) \]

Where \( A_L \) is the area under consideration, \( V \) is the volume, \( T_L \) is the temperature of the water volume and the subscripts 1 and 2 are the values at the beginning and end of \( \Delta t \).
Example: Evaporation from Lake Superior, 1975-1977

The annual cycle of air temperature, humidity, wind speed and evaporation over Lake Superior, 1975-1977. Note how in such large-deep lakes, heat storage (as expressed in terms of water temperature) strongly influences evaporation. Storage reaches a maximum in early autumn; this heat is then released upwards. This storage affect shifts the timing of peak evaporation with respect to the annual cycles of air temperature, water temperature and humidity.
The Penman equation (or combination equation)

The Penman equation combines mass transfer and the energy budget:

\[ E = \left[ \Delta \times (K + L) + \gamma \times K_E \times \rho_w \times \lambda_v \times v_a \times e_a^* \times (1 - W_a) \right] / \left[ \rho_w \times \lambda_v \times (\Delta + \gamma) \right] \]

Where \( K \) is net shortwave radiation, \( L \) is net longwave radiation, \( \gamma \) is the psychometric constant, \( K_E \) is the coefficient for the efficiency of the vertical transport of water vapor by turbulence, \( \rho_w \) is the density of water, \( \lambda_v \) is the latent heat of vaporization, \( v_a \) is the horizontal winds speed, \( e_a^* \) is the saturation vapor pressure of the air with respect to its temperature and \( W_a \) is the relative humidity of the air.

Note that \( \Delta = \frac{de^*}{dT} \) is the slope of the relation between saturation vapor pressure and temperature \( T \).

While the Penman equation it may look formidable, it essentially boils down to:

\[ E = (\Delta \times \text{Net radiation} + \gamma \times \text{mass transfer}) / (\Delta + \gamma) \]

The Penman equation assumes no ground heat flux, and no change in heat storage.
The Penman equation (cont.)

\[ E = \left[ \Delta(K+L) + \gamma \cdot K_E \cdot \rho_w \cdot \lambda_v \cdot v_a \cdot e_a^* \cdot (1-W_a) \right] / \left[ \rho_w \cdot \lambda_v \cdot (\Delta + \gamma) \right] \]

Just like the energy budget approach, it can be a challenge to directly measure all of the terms, typically requiring the use of empirical approximations for the radiation terms (this can be done for humidity as well). The Penman equation assumes no ground heat flux, and no change in heat storage.

Kohler and Parmele (1967) developed a method to include effects of water advected energy and changes in heat storage to provide a more generalized estimate of open water evaporation:

\[ E_L = E_P + \alpha_{KP} \cdot (A_w \cdot \Delta Q/\Delta t) \]

In which \( E_L \) is lake evaporation, \( E_P \) is the Penman evaporation, and \( \alpha_{KP} \) is the fraction of the net addition of energy from advection \( A_w \) and storage change \( \Delta Q/\Delta t \) that was used in evaporation over the interval \( \Delta t \).
Pan evaporation

A very direct approach to estimating free-water evaporation is to simply measure the changes in the depth or volume of water in a pan, also accounting for precipitation.

\[ E = P - (D_2 - D_1) \]

Where \( P \) is the precipitation over a given time and \( D_2 \) and \( D_1 \) are the depths at times 2 and 1 (this assumes a pan of uniform cross section so that volume change is directly proportional to depth change)

Issue: Pan evaporation will differ from evaporation over a lake because of the much smaller heat capacity of the water in the pan. Over the U.S., the average annual ratio of lake to pan evaporation is about 0.7. This is called the pan coefficient.

http://weather.uwaterloo.ca/info.html
Pan evaporation over the contiguous U.S.

The pattern of pan evaporation (here in archaic units of inches per year) across the U.S. is determined by latitude, continentality, elevation and average atmospheric circulation. This reflects controls by the availability of water at the surface, dryness of prevailing airmasses, energy input to the surface and turbulence.

Dingman 2002, Figure 7-7
Bare soil evaporation

Much of the earth’s surface has little or no vegetation cover, meaning little transpiration. Agricultural lands also have bare soil for much of the time. Bare soil evaporation is hence globally significant. Water at the surface is ranges from unlimited to limited. Water advected energy is zero and stored energy use is negligible.

Estimation strategies include:

1) Mass transfer
2) Energy budget
3) Penman approach
Bare soil evaporation

Following infiltration from rain, snowmelt or irrigation, evaporation from bare soil generally follows two stages:

1) Atmospheric controlled: The evaporation rate is largely determined by the surface energy balance and mass transfer conditions (wind and humidity) and is largely independent of soil moisture. Evaporation occurs close to the rate of free-water evaporation.

2) Soil controlled: The evaporation rate is determined by the rate at which water can be conducted to the surface in response to the suction due to an upward decrease in soil moisture. Soil characteristics are hence important. Evaporation is less than the free-water evaporation rate, the Bowen ration rises as the soil dries.

The transition between the two states is typically abrupt.
Transpiration

In transpiration, availability of water to the surface is limited, there is no water advected energy and stored energy use is negligible. Estimation methods include mass transfer approaches and the Penman-Monteith model.

**Key point:** Transpiration is a physical (not a metabolic) process. Movement of water vapor to the air from the stomata is replaced by evaporation from the walls of the stomata cavities; the loss of water induces movement of replacement water up through the plant’s vascular system. This produces a water content gradient between the roots and the soil, which induces movement of soil water into the roots, depleting the soil water and (via Darcy’s Law) drawing in soil water from surrounding regions. Roots in turn grow towards the soil water.

**Key point:** Plant can exert some control over the size of stomatal openings and hence these ease of vapor movement by the action of guard cells.

Dingman 2002, Figure 7-9
Size of stomatal openings

The figure at right shows a pair of guard cells from a poplar tree. Major factors influencing the opening or closing of guard cells are:

1) light: Most plants open their stomata during the day (for photosynthesis) and close them at night;
2) Humidity: Stomatal openings tend to decrease as humidity decreases below its saturation value;
3) The water content of leaf cells: If daytime leaf water contents get too low, stomata tend to close.

http://sols.unlv.edu/Schulte/Anatomy/Leaves/Leaves.html
Atmospheric and leaf resistance

Transpiration is a two-step process, in which water molecules must first pass from the stomatal cavities to the leaf surface, and the from the leaf surface to the atmosphere. One can think of this in terms of circuit analogy. Open water evaporation depends on the driving vapor pressure difference between the evaporating surface and the atmosphere, here represented by $\Delta e_v$ (the “voltage”) and an atmospheric resistance $1/C_{at}$, where $C_{at}$, the atmospheric conductance, represents the efficiency of turbulent eddies in transporting water vapor from the surface to the unsaturated air above. Transpiration includes an additional leaf resistance factor $1/C_{leaf}$, where $C_{leaf}$ is leaf conductance.
Atmospheric conductance

See Dingman et al. (2002) for estimation of atmospheric conductance $C_{at}$ (equation 7-47 to 7-51). Key points: Conductance increases with the horizontal wind speed $v_a$ and for a given wind speed the conductance is higher for higher vegetation types. The latter relates to the surface roughness; the higher the vegetation, the greater is the roughness, the more turbulent is the flow, and the more efficient the upward transfer of water vapor.
Leaf conductance

Maximum leaf conductance depends on the number of stoma per unit area (the stomatal density) and the size of the stomatal openings, which depends on the species. As noted, plants in turn can control the size of stomatal openings, and hence the leaf conductance. Following Stewart (1988) leaf conductance can be expressed as follows:

$$C_{\text{leaf}} = C^*_{\text{leaf}} \cdot f_k(K_{\text{in}}) \cdot f_p(\Delta \rho_v) \cdot f_T(T_a) \cdot f_\theta(\Delta \theta)$$

Where $C^*_{\text{leaf}}$ is the maximum leaf conductance, $K_{\text{in}}$ is the incident solar radiation, $\Delta \rho_v$ is the humidity deficit (the difference between the saturated and actual humidity of the air), $T_a$ is air temperature, $\Delta \theta$ is the soil moisture deficit, and the “$f$” coefficients represent the effects of each environmental factor on the leaf conductance they are non-linear coefficients with values ranging from 0 to 1 (derived from observations; their form seems to be fairly universal).
The “f” coefficients in the previous equation for leaf conductance are highly non-linear. Note, for example, how the coefficient for solar radiation initially rises sharply with increasing solar radiation, but then flattens out. As another example, leaf conductance increases with increasing air temperature up to about 18°C, but then decreases at higher temperatures.

Dingman 2002, Figure 7-13
Leaf conductance (cont.)

The table at right shows typical values of maximum leaf conductance, leaf area index (LAI), albedo $\alpha$ and vegetation height $z_{veg}$.

$$\text{LAI} = \frac{LS}{A}, \text{ where LS is the total area of leaf surface above ground area A.}$$

This leads to the concept of canopy conductance:

$$C_{can} = f_s \cdot \text{LAI} \cdot C_{\text{leaf}}$$

Where $f_s$ is a shelter factor, which accounts for some of the leaves being sheltered from the sun and wind and which hence transpire at lower levels than unshaded leaves exposed to the full wind.

<table>
<thead>
<tr>
<th>Land Cover</th>
<th>$C_{\text{leaf}}$ (mm s$^{-1}$)</th>
<th>LAI</th>
<th>$\alpha$</th>
<th>$z_{veg}$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Conifer forest</td>
<td>5.3</td>
<td>6.0</td>
<td>0.14</td>
<td>25.0</td>
</tr>
<tr>
<td>Broadleaf forest</td>
<td>5.3</td>
<td>6.0</td>
<td>0.18</td>
<td>25.0</td>
</tr>
<tr>
<td>Savannah/shrub</td>
<td>5.3</td>
<td>3.0</td>
<td>0.18</td>
<td>8.0</td>
</tr>
<tr>
<td>Grassland</td>
<td>8.0</td>
<td>3.0</td>
<td>0.20</td>
<td>0.5</td>
</tr>
<tr>
<td>Tundra/nonforest</td>
<td>6.6</td>
<td>4.0</td>
<td>0.20</td>
<td>0.3</td>
</tr>
<tr>
<td>Desert</td>
<td>5.0</td>
<td>1.0</td>
<td>0.26</td>
<td>0.1</td>
</tr>
<tr>
<td>Typical crop</td>
<td>11.0</td>
<td>3.0</td>
<td>0.22</td>
<td>0.3</td>
</tr>
</tbody>
</table>

Data from Federer et al. (1996).

Dingman 2002, Table 7-5
The Penman-Monteith equation

Monteith (1965) showed how the Penman equation can be modified to represent the ET rate from a vegetated surface by incorporating atmospheric conductance and canopy conductance:

\[ E = \frac{[(K+L) + \rho_a \cdot c_a \cdot C_{at} \cdot e_a^* \cdot (1-W_a)]}{\rho_w \cdot \lambda_v \cdot (\Delta + \gamma \cdot (1 + C_{at}/C_{can}))} \]

Where \( K \) is net shortwave radiation, \( L \) is net longwave radiation, \( \gamma \) is the psychometric constant, \( \rho_w \) is the density of water, \( \lambda_v \) is the latent heat of vaporization, \( C_{at} \) is the atmospheric conductance, \( C_{can} \) is the canopy conductance, \( e_a^* \) is the saturation vapor pressure of the air with respect to its temperature and \( W_a \) is the relative humidity of the air.

Like the Penman equation, the Penman-Monteith equation assumes no ground heat flux, and no change in heat storage.
Canopy interception

Canopy interception refers to the process by which precipitation falls on a vegetative surface (a canopy). Intercepted water is then evaporated; this can be a significant fraction of total ET in most regions. Interception depends on vegetative type and stage of development (linked to leaf area index) and the intensity, duration and form of precipitation. Key definitions used in describing and measuring canopy interception are gross rainfall ($R$), throughfall ($R_t$), stemflow ($R_s$, water that flows down trunks and stems), canopy interception loss ($E_c$, water that evaporates from the canopy), litter interception loss $E_l$ (water evaporating from near ground plants and leaf litter), total interception loss ($E_i = E_c + E_l$)) and net rainfall ($R_n = R - E_i$)
Potential evapotranspiration (PET)

PET is the rate at which ET would occur from a large area completely and uniformly covered with growing vegetation with access to an unlimited supply of soil water and without advection and heat storage effects. The concept was originally designed for use in climate classification by Thornthwaite (1948).

However, we know that ET depends on vegetative characteristics, such as

• Albedo of the vegetative surface
• Maximum leaf conductance
• Atmospheric conductance, which depends on vegetation height
• The presence or absence of intercepted water

Note that PET will vary with respect to available energy; change the available energy and PET changes.

Penman (1956) redefined PET as water transpired “by a short green crop, completely shading the ground, of uniform height and never short of water”

“Reference-crop ET” is a synonym for PET.
PET (cont.)

Many methods have been offered; in practice PET depends on the method employed. These include:

1) Temperature-based approaches, for example, the empirical equation of Harmon (1963) is:

\[ \text{PET} = 29.8 \times D \times \left( \frac{e^*}{a} \right) / (T_a - 273.2) \]

Where \( D \) is daylength in hours, and \( e^* \) is the saturation vapor pressure at the mean daily temperature \( T_d \).

2) Radiation based methods, such as from Priestly and Taylor (1972):

\[ \text{PET} = \left( \alpha_pT \times (\Delta \times (K+L)) \right) / (\rho_w \lambda_v (\Delta + \gamma)) \]

This is termed equilibrium PET, were \( \alpha_pT = 1.26 \) is empirically derived and has been shown to apply to a wide range of well-watered surfaces.

3) Some version of the Penman equation

4) Pan evaporation (PET over short vegetation is commonly very similar to free-water evaporation).
Actual ET: The ET that actually occurs

Putting together all that has been covered so far, it is clear that actual ET depends on many factors, including:

• Whether water is freely available at the surface (e.g., over a lake, wetland or well-watered crop) or is limited

• The humidity of the overlying air

• The nature of turbulent eddies

• How much energy is available to break hydrogen bonds

• The type of vegetation, which determines stomatal size and density

• Environmental factors influencing the size of stomatal openings

A couple of generalities:

• In hot, arid regions, PET greatly exceeds precipitation so that actual ET is water limited, and roughly equal to precipitation.

• In regions with abundant precipitation in all season, ET is limited by available energy, so actual ET is essentially equal to average PET
Lysimeters are artificially enclosed volumes of soil including any vegetation growing on top for which inflows and outflows of water can be measured, and, commonly, changes in storage can be measured by weighing. They can range from the very small and simple (1 m³ or less) to the very large (150 m³). The vegetation growing from the enclosed soil volume is intended to be representative of the surrounding region. Data from well-functioning lysimeters are often viewed as providing the “true” actual ET. The lysimeter depicted at right is a fancy one.
The water balance approach for estimating actual ET for a watershed

Inputs (I), outputs (O) and storage (S):
I: Precipitation (P)
   Groundwater in (G_{in})
O: Evapotranspiration (ET)
   Groundwater out (G_{out})
   River discharge (Q)
Storage (S): In groundwater, rivers and lakes

\[ \Delta S = P + G_{in} - (Q + ET + G_{out}) \]

If we assume that \( G_{in} \) and \( G_{out} \) are negligible, and that for the long-term annual mean, \( \Delta S \) is zero, then:

\[ P = ET + Q, \text{ or } ET = P - Q \]
Middle Boulder Creek: 06725500 (Nederland)

Average Q = 54.4 cfs (ft$^3$ s$^{-1}$)  
= 1.54 cms (m$^3$ s$^{-1}$)  
= 4.86x$10^7$ m$^3$ yr$^{-1}$

Drainage area = 36.2 mi$^2$  
= 93.8 km$^2$  
= 9.38x$10^7$ m$^2$

$R = \frac{4.86x10^7 \text{ m}^3 \text{ yr}^{-1}}{9.38x10^7 \text{ m}^2} = 52 \text{ cm yr}^{-1}$

Annual Precipitation (P):
Berthoud Pass = 95 cm
Niwot Ridge: 93 cm
Gross Reservoir 53 cm
City of Boulder: 48 cm (19 in)

If we assume that the precipitation at Berthoud pass is representative of the Middle Boulder Creek watershed, then

$ET = Q - R$

$ET = 95 \text{ cm} - 52 \text{ cm} = 43 \text{ cm}.$

Obvious question: Is the precipitation from a single station a reasonable approximation of the watershed-average precipitation? Likely not.

http://czo.colorado.edu/html/sites.shtml
Mean actual ET for the Arctic terrestrial drainage (mm) for alternate months, estimated as a residual from bias-adjusted precipitation data and aerological estimates of P-ET from the NCEP/NCAR reanalysis. Results are based on data from 1960 through 1999 [from Serreze et al., 2003a, by permission of AGU]. Actual ET peaks in July. While this residual approach can provide to useful estimates of ET, the value obtained reflects the errors in both P and in P-ET; one manifestation of these errors is regions of strongly negative ET in winter.
Annual mean actual ET over the Arctic terrestrial drainage from different land surface models

Compiled by D. Slater