Advanced Agro-Hydro- Meteorology

A MSc course for students of Atmospheric Sciences

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Lecture 4: Evaporation and transpiration

4.1 Introduction

- Evaporation is the process by which a liquid turns into a gas. It is also one of the three main steps in the global water cycle.
- For evaporation to occur it is necessary to have:
  (i) a supply of water; (ii) a source of heat such as direct solar energy R_c, sensible heat H, heat from the ground G, or stored heat from the water R_s; and (iii) a gradient of concentration $e_s – e_d$, where e_s is the saturated vapor pressure at temperature T and e_d is the vapor pressure for dry air.
- The physics of evaporation is the same regardless of the evaporating surface, although different surfaces, such as open water, bare soils and vegetation covers, impose different controls on the processes.
- Sometimes total evaporation is referred to as evapotranspiration (the sum of evaporation from the land surface plus transpiration from plants); and transpiration (evaporation of water from plant tissue) from plants, which is closely linked to photosynthesis.
- *Potential evaporation* is defined as the amount of water that could be evaporated were it available. It is a function of surface and air temperatures, insolation, and wind.

4.2 Modelling potential evaporation based upon observations

Monthly potential evaporation $E_p(cm)$ is calculated as an exponential function of air temperature:

$$E_p = \frac{16t}{I} \qquad (4.1)$$

where t (°C) is the mean monthly temperature and I is the annual heat index (heat index is an index that combines air temperature and relative humidity), the sum of 12 individual monthly indices i, where

$$i = (\frac{t}{5})^{1.514} \qquad (4.2)$$

The formula seems to work well in the temperate, continental climate where temperature and radiation are strongly correlated (Such as USA). The heat index can be calculated from the following equation (T is Temp. & R is RH):

Heat Index = $-42.379 + 2.04901523T + 10.14333127R - 0.22475541TR - 6.83783 \times 10^{-3}T^2 - 5.481717 \times 10^{-2}R^2 + 1.22874 \times 10^{-3}T^2R + 8.5282 \times 10^{-4}TR^2 - 1.99 \times 10^{-6}T^2R^2$

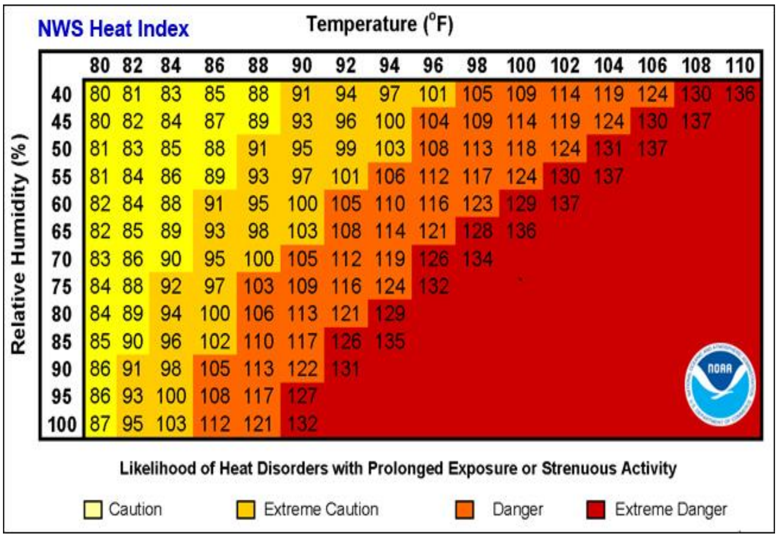
What is the heat index?

"It's not the heat, it's the humidity". That's a partly valid phrase you may have heard in the summer, but it's actually both. The heat index, also known as the apparent temperature, is what the temperature feels like to the human body when relative humidity is combined with the air temperature. This has important considerations for the human body's comfort. When the body gets too hot, it begins to perspire or sweat to cool itself off. If the perspiration is not able to evaporate, the body cannot regulate its temperature. Evaporation is a cooling process. When perspiration is evaporated off the body, it effectively reduces the body's temperature. When the atmospheric moisture content (i.e. relative humidity) is high, the rate of evaporation from the body decreases. In other words, the human body feels warmer in humid conditions. The opposite is true when the relative humidity decreases because the rate of perspiration increases. The body actually feels cooler in arid conditions. There is direct relationship between the air temperature and relative humidity and the heat index, meaning as the air temperature and relative humidity increase (decrease), the heat index increases (decreases).

Example: Using the chart determine the heat index if the air temperature is 88°F and the relative humidity is 80%. Then determine the heat index for same temperature but for 40% humidity.

Solution: for T= 88 °F and RH= 80% the heat index i=106 °F Danger (≈41 °C)
for T= 88 °F and RH= 40% i=88 °F Caution (≈31 °C)

[Note: F= C×1.8 +32
and C= (F-32)×5/9]


NWS Heat Index

Relative Humidity (%) \ Temperature (°F)	80	82	84	86	88	90	92	94	96	98	100	102	104	106	108	110
40	80	81	83	85	88	91	94	97	101	105	109	114	119	124	130	136
45	80	82	84	87	89	93	96	100	104	109	114	119	124	130	137	
50	81	83	85	88	91	95	99	103	108	113	118	124	131	137		
55	81	84	86	89	93	97	101	106	112	117	124	130	137			
60	82	84	88	91	95	100	105	110	116	123	129	137				
65	82	85	89	93	98	103	108	114	121	128	136					
70	83	86	90	95	100	105	112	119	126	134						
75	84	88	92	97	103	109	116	124	132							
80	84	89	94	100	106	113	121	129								
85	85	90	96	102	110	117	126	135								
90	86	91	98	105	113	122	131									
95	86	93	100	108	117	127										
100	87	95	103	112	121	132										

Likelihood of Heat Disorders with Prolonged Exposure or Strenuous Activity

Caution Extreme Caution Danger Extreme Danger

4.3 Aerodynamic approach

- Dalton (1801) provided a means of quantifying evaporation rates using an aerodynamic approach. The equation for evaporation E, sometimes referred to as the *Dalton equation is*:

$$E = (e_0 - e)\ f(u) \qquad (4.3)$$

where e_0 is the vapor pressure at the surface, e is the vapor pressure of the air, and *f(u)* is a function of wind speed.

The aerodynamic function is expressed as an *aerodynamic resistance* r_a to the transfer of water vapor down the vapor concentration gradient that exists between the evaporating surface and the atmosphere, as shown in Figure 4.1.

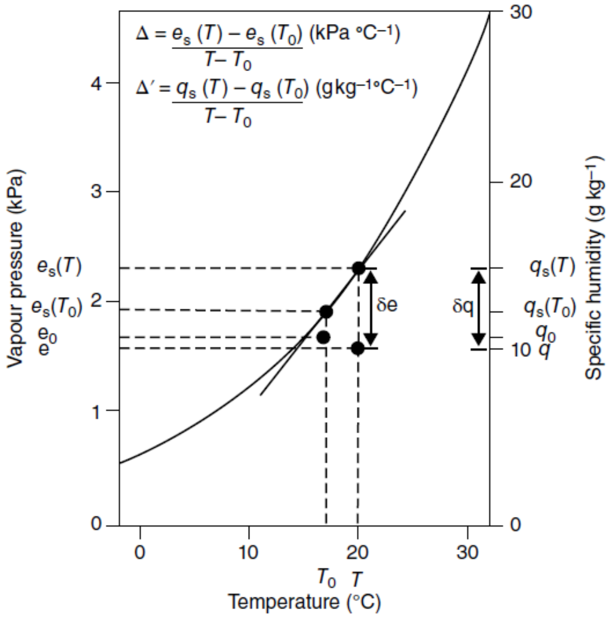


Figure 4.1

***Aerodynamic resistance**: (Also called drag* or aerodynamic drag.) The component of force exerted by the air on a liquid or solid object (such as a raindrop or airplane) that is parallel and opposite to the direction of flow relative to the object.

The *evaporation rate* is expressed as the equivalent rate of latent heat λE, where λ is the latent heat of vaporization of water, so that

$$\lambda E = \frac{\lambda \rho_a \varepsilon}{r_a}\left(\frac{e_0}{p - e_0} - \frac{e}{p - e}\right) \qquad (4.4)$$

where ρ_a is the density of dry air, ε is he ratio of the molecular weights of water and air, and p is the total air pressure; then equation 4.4 can be expressed as

$$\lambda E = \frac{\rho C_p (e_0 - e)}{r_a \gamma} \qquad (4.5)$$

where C_p is the specific heat of air and $\gamma = C_p p / \lambda \varepsilon$ is known as the *psychrometric constant*, which has a value of 0.066 kPa °C^{-1} at a temperature of 20 °C and a pressure of 100 kPa. Note that γ is not a constant, although it is referred to as such, because it varies with atmospheric pressure and temperature.

Exercise: Estimate evaporation rate assuming the specific heat of air is $C_p = 1.0 \times 10^3\ J\ kg^{-1}\ K^{-1}$ at a temperature of 20 °C, with air pressure = 100 kPa, vapour pressure = 2.3 kPa, air density = 0.00129 g cm^{-3} and aerodynamic function for a rough surface = 3.5 s m^{-1}.

Sol. Specific humidity q may be defined as the mass of water vapor contained in a unit mass of moist air (kg kg^{-1}; g kg^{-1}); then

$$q = \frac{\varepsilon e}{p - (1 - \varepsilon) e} \qquad (4.6)$$

Equation 4.5 may then be rewritten as

$$\lambda E = \frac{\lambda \rho (q_0 - q)}{r_a} \qquad (4.7)$$

where ρ is the density of moist air andq_0 and q are the specific humidities of the surface and the air respectively.

4.4 Energy balance

The energy balance approach to estimating evaporation involves the process of vapor transfer, the factors in which are the following:

R_c: the incoming radiation from the surface (reflected short wave or long wave)
$R_c(1\text{–}alb)$: the incoming radiation into a surface of albedo (*alb*)

R_b: the outgoing radiation from the surface, which may be reflected short wave or long wave
H: the sensible heat transfer from air to surface or in the opposite direction. The sensible heat *H* may be given as

$$H = \frac{\rho\, C_p(T_0 - T)}{r_a} \qquad (4.8)$$

where T_0 and T are the surface and air temperatures respectively.

LE: the heat used in converting liquid to vapor, where L is the latent heat and E is the evaporation

G: the heat flux into the ground or vegetation or in the opposite direction

R_s: heat stored in the water

R_p: heat converted to chemical energy in the process of photosynthesis

R_i: heat moved into the air or out of the system by water inflow or outflow

R_n: the net radiation received by the surface, where $R_n = R_c(1 - alb) - R_b$.

Hence at the evaporating surface, the conservation of energy gives

$$R_n = H + LE + G + R_s + R_p + R_i \qquad (4.9)$$

Neglecting the storage terms, which are not usually significant over short periods, gives

$$R_n = H + LE + G \qquad (4.10)$$

By combining the aerodynamic and energy balance methods, we can compute the evaporation.

4.5 The Penman equation

The Penman equation describes evaporation (E) from an open water surface, and was developed by Howard Penman in 1948. Penman's equation requires daily mean temperature, wind speed, air pressure, and solar radiation to predict E. The Penman–Monteith equation approximates net evapotranspiration (ET) from meteorological data, as a replacement for direct measurement of evapotranspiration. The equation is widely used, and is the standard method for modeling evapotranspiration used by the United Nations FAO

In practice, it is not usually possible to solve either the aerodynamic or the energy balance equations, as the surface temperature, the humidity terms, and the sensible heat term are generally unknown. However, Penman (1948) provided a solution using knowledge of the *change with temperature of the saturated vapor pressure of water*.

In Figure 4.1,

$$\Delta = \frac{e_s(T) - e_s(T_0)}{T - T_0} \qquad (4.11)$$

where e_s represents the saturated vapour pressure at the temperature T or T_0. If the surface of the vegetation is wet, the humidity at the surface expressed as a vapor pressure is given by

$$e_0 = e_s(T_0) \qquad (4.12)$$

Hence eliminating surface temperature, surface humidity and sensible heat from equations 4.7, 4.8, 4.11 and 4.12 gives

$$\lambda E = \frac{\Delta H + \dfrac{\rho C_p (e_s(T) - e)}{r_a}}{\Delta + \gamma} \qquad (4.13)$$

Subsequently several developments of the Penman equation have been derived to allow for the empirical aerodynamic and net radiation terms.

4.6 Sensible and water vapor fluxes

Eddy diffusion theory (flux–gradient theory) provides a model for the transport of momentum, heat and water vapor between the surface and the atmosphere. Transfer coefficients for the various entities may be calculated.

Bowen (1926) showed that heat and water transfer, assuming laminar flow, are proportional: hence the *Bowen ratio*

$$\beta = \frac{H}{LE} \qquad (4.17)$$

This means that there is a constant division of the available energy between the transfer of heat and water vapor. The Bowen ratio may be given by

$$\beta = \frac{\rho C_p K_H \dfrac{\partial T}{\partial z}}{L K_W \dfrac{\partial q}{\partial z}} \qquad (4.18)$$

where K_H is the transfer coefficient of heat, K_W is the transfer coefficient of water vapor, C_p is the specific heat at constant pressure, and the psychrometric constant is $\gamma = \frac{C_p p}{0.622\lambda} = 0.0677 kPa\,°C^{-1}$, where p is the total air pressure and λ is the latent heat of vaporization of water at $20\ °C$ and a pressure of $101.2\ kPa$. Δq and ΔT are the specific humidity and temperature differences measured over the same height interval. However, in convective conditions Bowen's ratio and the ratio of K_H and K_W can vary considerably.