

Lecture 15, Water, Humidity, Pressure and Trace Gases, Part 2

October 6, 2014

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Topics to be covered:

1. Atmospheric measures of humidity
2. Observations of humidity
 1. diurnal and seasonal patterns
 2. vertical gradients

Vapor Pressure Deficit:

The saturation vapor pressure deficit is an important variable in biometeorology, as it defines the potential for mass transfer. The vapor pressure deficit is defined below and is related to relative humidity:

$$D = e_s(T) - e_a = e_s(T)(1 - h_r)$$

Over the course of a day the vapor pressure deficit is maximal late in the day, as it coincides with the time of maximum temperature and e_s .

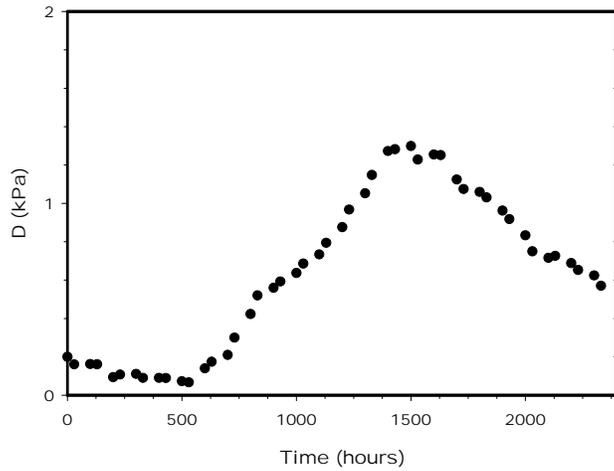
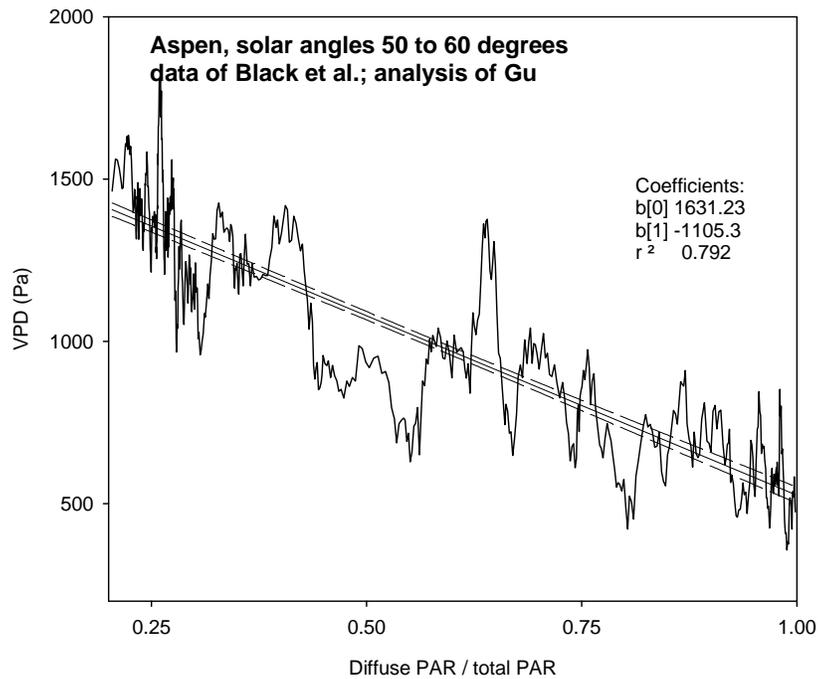


Figure 1 Diurnal course of vapor pressure deficit over a wheat crop in eastern Oregon

Typically vpd is low under cloudy days and high under clear skies. Quantification of this effect is shown below



Slope of the saturation vapor pressure-temperature relation, $de_s(T)/dT$ (kPa °C⁻¹)

Vapor pressure deficit is an important variable for computing evaporation, as via the the Penman-Monteith Equation. An intrinsic part of this equation is the linearization of the difference between the saturation vapor pressure at the surface temperature and that of the free air. This relation reduces to an algebraic equation that includes D and the slope of the saturation vapor pressure-temperature relation:

$$e_s(T_s) - e_a = (e_s(T_a) - e_a) - s(T_a - T_s)$$

To evaluate the Penman-Monteith equation correctly, we need to evaluate the **slope of the saturation vapor pressure**.

$$\frac{de_s(T)}{dT} = s = \frac{17.502 \cdot 240.97 \cdot e_s(T)}{(240.97 + T)^2}$$

The following figure shows that s is a strong function of temperature

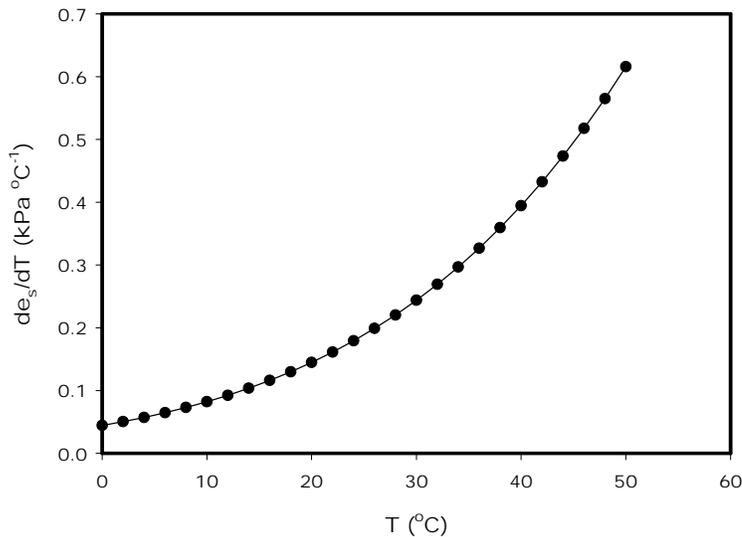


Figure 2 Functional relation between slope of the saturation vapor pressure curve and temperature

Dew point temperature

What is the temperature at which dew forms? This temperature when the saturation vapor pressure equals the actual vapor pressure or the relative humidity equals one.

$$T_d = \frac{240.97 \cdot \ln(e_a / 0.611)}{17.502 - \ln(e_a / 0.611)}$$

Vapor pressure has units of kPa, dew point temperature is in degrees C.

The relationship between T_{dew} and vapor pressure is:

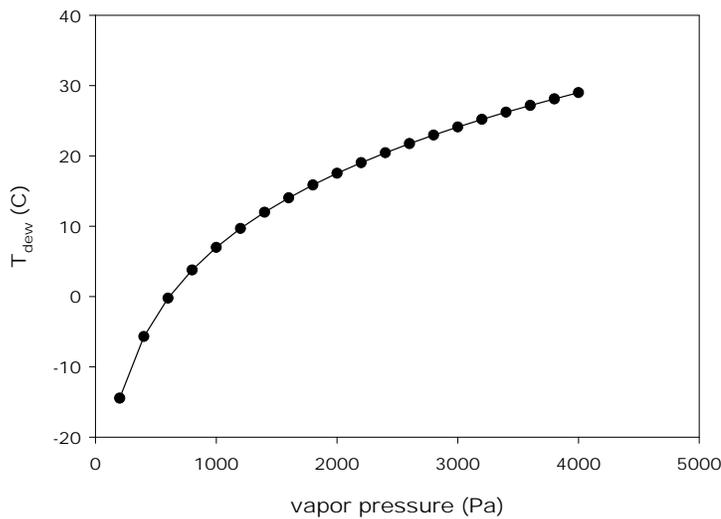


Figure 3 The relation between T_{dew} and vapor pressure

Interested in fabricating a cheap humidity sensor or impressing your friend by telling them what the humidity of the air is? Take a glass of water and interest a thermometer and ice in it. Watch it and determine what temperature dew forms on the outside of the glass.

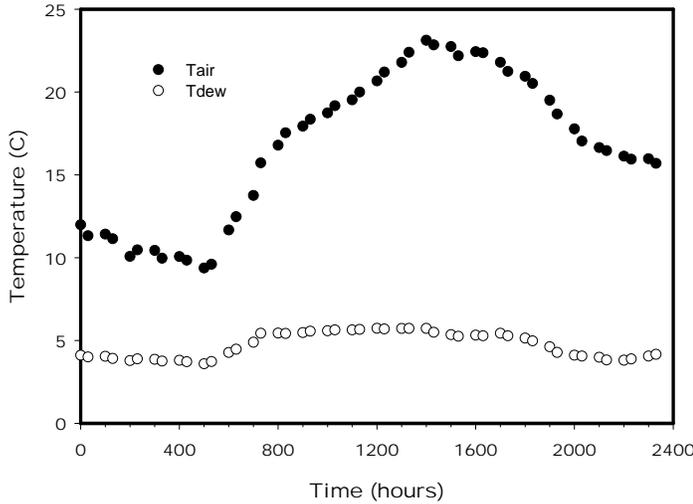


Figure 4 Diurnal pattern of air and dew temperature

H. Wet Bulb Temperature and Psychrometric relation.

Ever notice how the air temperature drops as you drive past an irrigated alfalfa field at night? Ever wonder how swamp coolers on western homes work? They use the wet bulb principle to cool the house. In practice, air blows over a wet surface until it drops to its wet bulb temperature.

The wet bulb temperature occurs from **adiabatic evaporation**, evaporation without a change in heat. The wet bulb temperature is defined as the temperature at which the change in the air's heat content equals the latent heat evaporated into the air. It represents the balance between latent and sensible heat exchange. In other words, it takes heat to evaporate water, but this heat is extracted from the air, so there is no net heat exchange, but the temperature of the air does lower and the humidity of the air increases.

Hence, **wet bulb psychrometry** is another simple way to determine the humidity of the air.

The wet bulb temperature is generally equal to or less than air temperature:

$$C_p(T_a - T_w) = \frac{\lambda(e_s(T_w) - e_a)}{P}$$

The humidity, vapor pressure can be assessed by algebraic manipulation of the previous equation.

$$e_a = e_s(T_w) - \frac{C_p}{\lambda} P(T_a - T_w)$$

A psychrometer is an instrument that consists of two thermometers, one dry and the other with a wet wick. In practice, the psychrometer must be properly ventilated and not exposed to direct sun. A flow velocity of at least 4 m s^{-1} is necessary. The term $\frac{C_p}{\lambda}$ is the psychrometric constant (γ). Its value is 0.661 kPa K^{-1} at 25 C and it ranges between 0.646 and 0.675 as temperature increases from 0 to $45 \text{ }^\circ\text{C}$.

Phase change, concept of Latent heat

Latent heat of evaporation is associated with the phase transition liquid to vapor. The energy involved is computed from:

$$\lambda = 3149000 - 2370 \cdot T_k \quad (\text{J kg}^{-1})$$

Its value is 2450 J g^{-1} at $20 \text{ }^\circ\text{C}$

Latent heat of fusion (phase transition from solid, ice, to vapor)

334 J g^{-1} at 0 C

Latent heat of sublimation (phase transition from solid to vapor)

2830 J g^{-1} at $^\circ\text{C}$

Humidity and the Plant Microclimate

We have seen so far that temperature, relative humidity, and vapor pressure deficit may swing widely over the course of a day, but the absolute humidity and vapor pressure are relatively constant. These measures tend to vary more over the course of the year as the supply of moisture changes. Warmer air can also hold more moisture. This effect can be demonstrated by plotting humidity versus temperature.

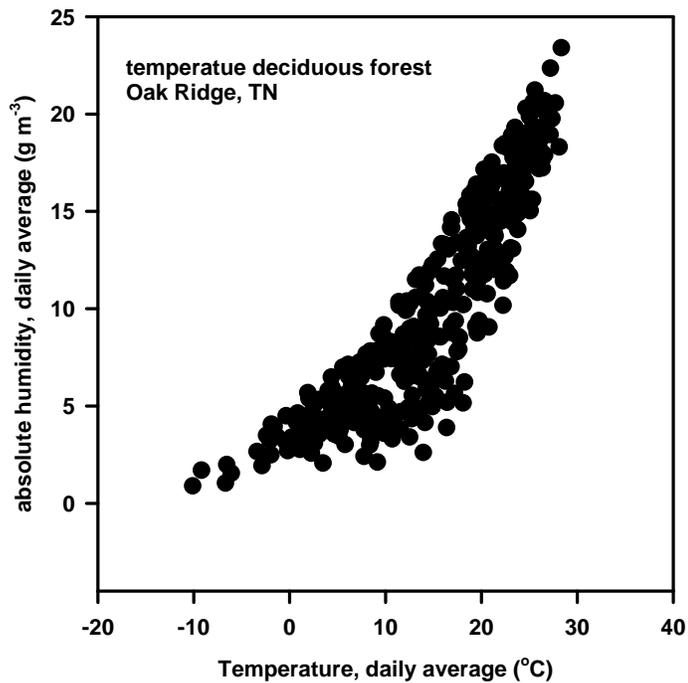


Figure 5 Relation between absolute humidity and average temperature. Warmer air can hold more moisture, hence the sensation felt in the humid tropics.

We can also deduce humidity by relating dew point temperature with the minimum temperature. One can expect a close relation between the two variables because radiative cooling can occur at night until condensation occurs, causing fog which would re-radiate energy and minimize further cooling, or a supply of energy being supplied by condensation, offsetting cooling. We explore this relation for two sites, one in the humid east and another in the semi-arid west. In both cases over 70% of the variance in minimum daily temperature explains variation in dew point temperature. The slope is greater than one for the east and less than one for the west.

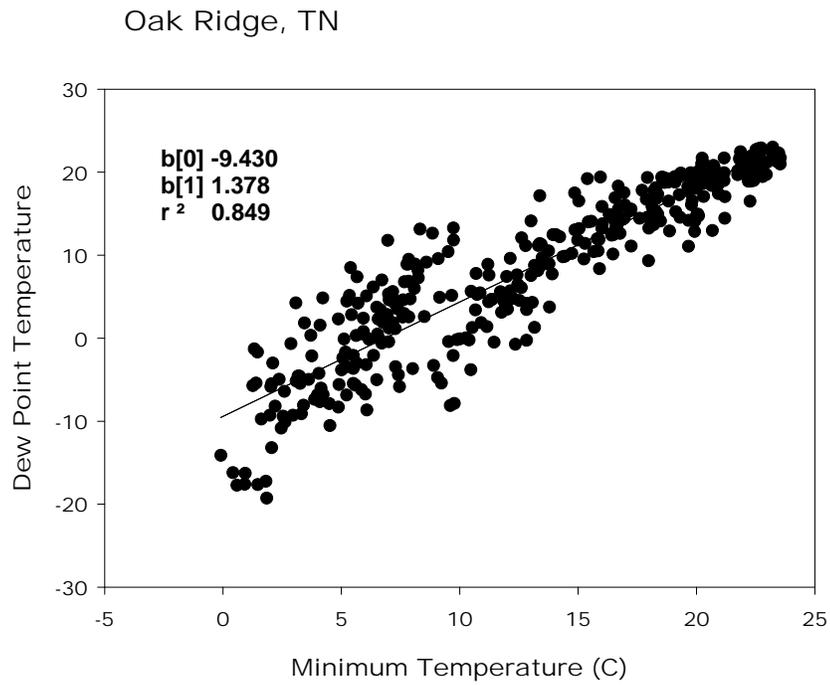


Figure 6 Relation between dew point and minimum temperature for a humid eastern location

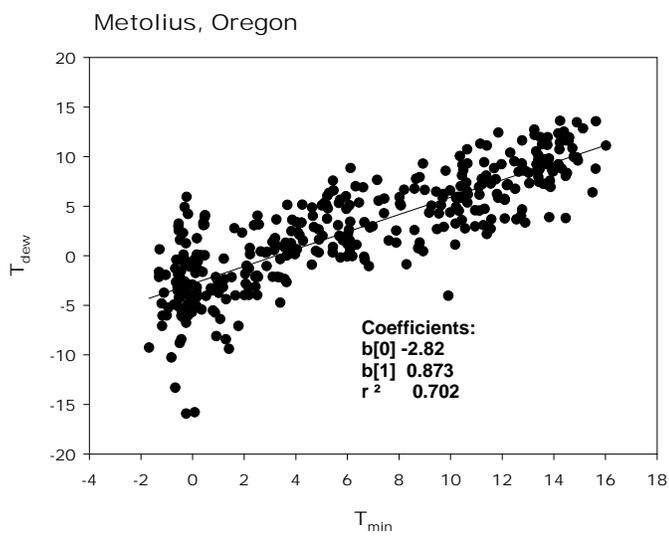


Figure 7 Relation between T_{dew} and minimum temperature for a semi-arid western location

Interactions between plant function and climate affect humidity deficits. Seasonal variations in humidity are shown for the case of a D over a deciduous forest in Massachusetts. The values of D are lowest during the winter when the forest is dormant and not transpiring and there is snow on ground. During the summer transpiration, vapor pressure deficit is large, reflecting the effect of warm daytime air temperature on the computation of the saturation vapor pressure.

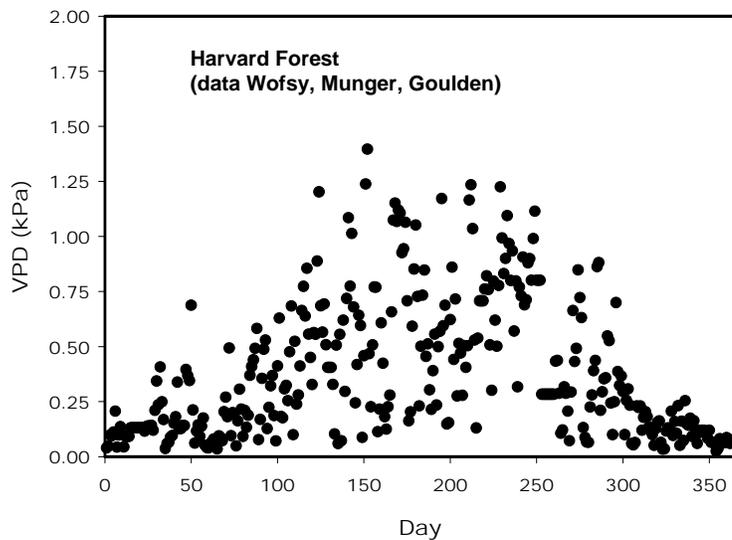


Figure 8 Seasonal variation in vapor pressure deficit over a forest in Massachusetts. (Data of Munger, Wofsy, Goulden)

In a seminal study by Schartz and Karl (Schwartz, Karl, 1990) they were able to demonstrate direct links between plant phenology and temperature. Humidity changes by a factor of three before and after leafout

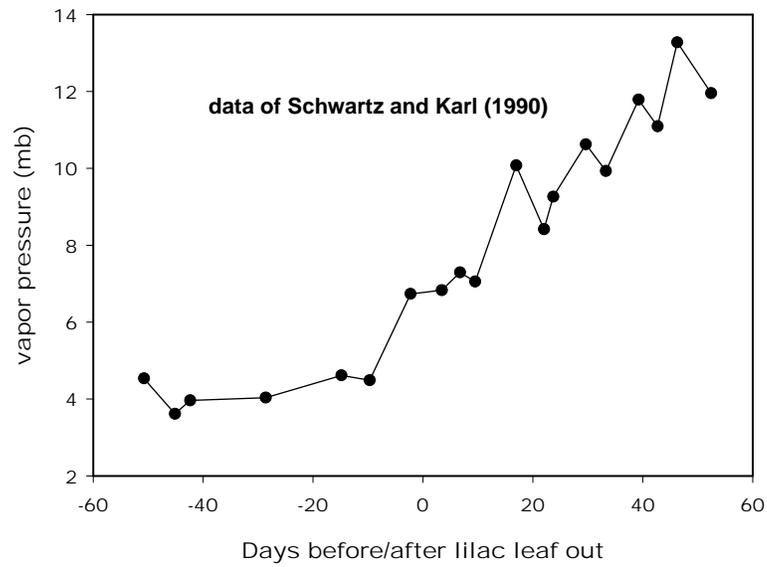


Figure 9 Data of Schwartz and Karl

We can look at the results for a given site, too. By looking at vapor pressure course at the Oak Ridge, TN site we see that the coolest month has lower q because low evaporation

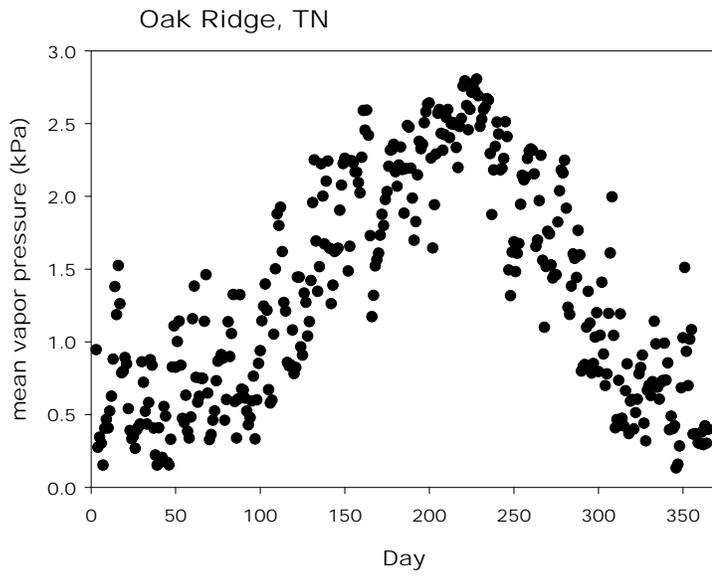


Figure 10 Mean vapor pressure. Annual course at Oak Ridge, TN. Notice the break in the curve around day 100, when leaf out occurs and a new source of water is introduced via plant transpiration.

Humidity profiles evolve as one passes from a dry to wet landscape or vice versa. We have made measurements across a desert-crop interface and used the advection model of Rao et al to quantify the spatial evolution of humidity. At the upwind edge of an irrigated crop the vertical profile is quite weak, as evaporation is low and the background air is very dry. As one moves farther into a field, one observes stronger gradients near the crop-air interface and a moistening of the bulk air layer. By the time an equilibrium profile is developed (800 m downstream), humidity has increased by 2 g m^{-3} near the surface and by about 1 g m^{-3} in the surface boundary layer.

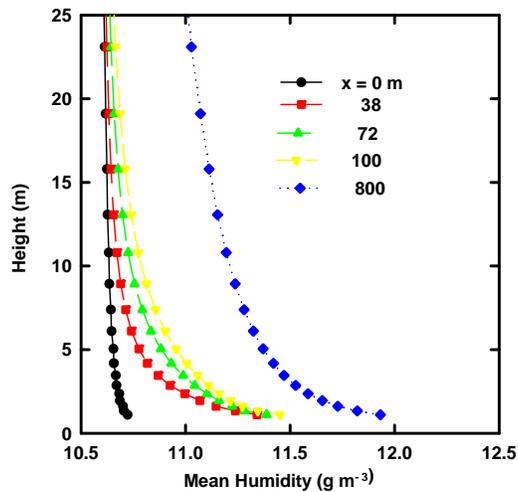


Figure 11 Evolution of humidity profiles across a transition of a dry desert and evaporating potato field (model calculations) (Baldocchi, Rao, 1995)

A biophysical model can be used to look at humidity and flux profiles in and about a forest for a wide range of conditions. Below we show computations during the day and night. During the day there is a kink in the profile, where leaf area density and evaporation is greatest. This is also a region of counter gradient transfer. Though in general gradients of humidity are weak in the canopy, despite significant rates of evaporation. At night the gradient profile is even weaker and a change in sign of the flux profile indicates dew deposition to the forest.

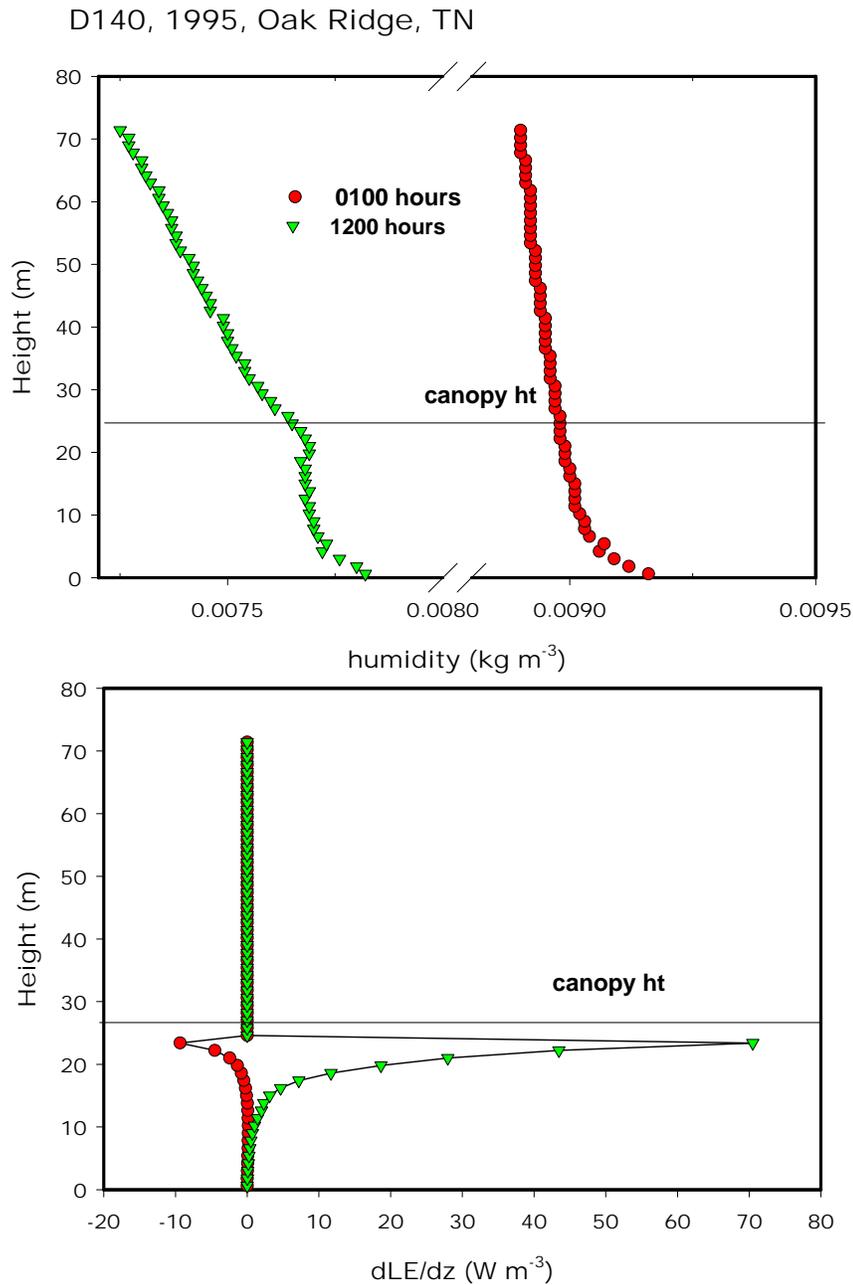


Figure 12 Calculations of vapor pressure and source profiles in a broadleaved forest for a daytime and night period.

It is fruitful to compare theory and reality. Below we show some profiles of humidity measured in and above a boreal aspen stand. Like the model computations, gradients of humidity are rather weak in and about the forest. Positive gradients during the day indicate evaporation and negative and inverted gradients at night infer dew deposition.

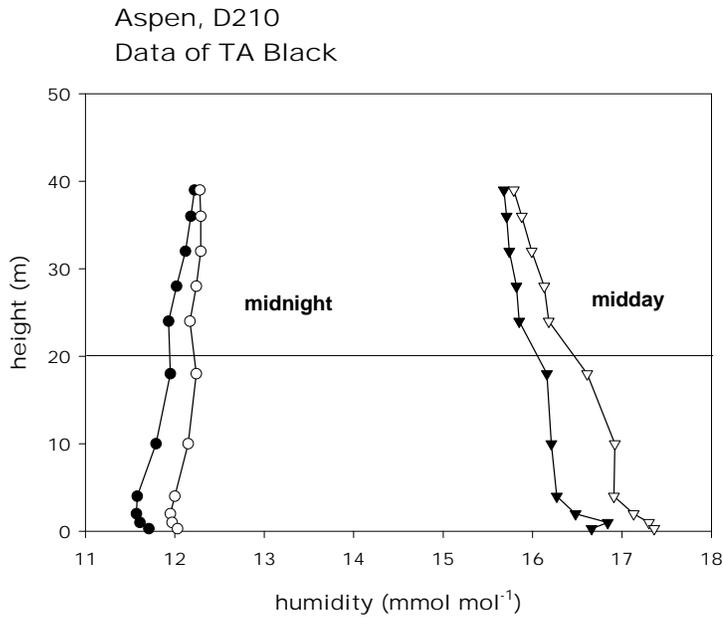


Figure 13 Humidity profiles within and above an aspen forest in Canada during the day and night. Data are from Dr. T.A. Black, UBC.

Pressure

From the gas laws, several useful relationships for pressure can be defined. If we are interested in measuring pressure fluctuations, which can be of importance for studying soil gas exchange, we must recognize links between temperature and pressure fluctuations. These can be quantified by taking the derivative of the gas law with respect to temperature.

$$\frac{dP}{dT} = \frac{\rho R}{m} = 344 \text{ PaK}^{-1}$$

Hydrostatic Equation

$$\frac{dp}{dz} = -g\rho$$

integration of the hydrostatic equation produces a relation showing how pressure varies with height.

$$p(z) = \int_z^{\infty} g \rho dz$$

$$\frac{dp}{p} = -\frac{g}{RT} dz$$

$$p = p(0) \exp\left(-\frac{g}{R} \int_0^z \frac{dz}{T}\right)$$

We observe that pressure decreases exponentially with height.

A time trace of pressure is shown below. Over the course of a year peak to peak differences in daily means are about 30 mb, for this continental site. The passage of a hurricane would extend this range markedly.

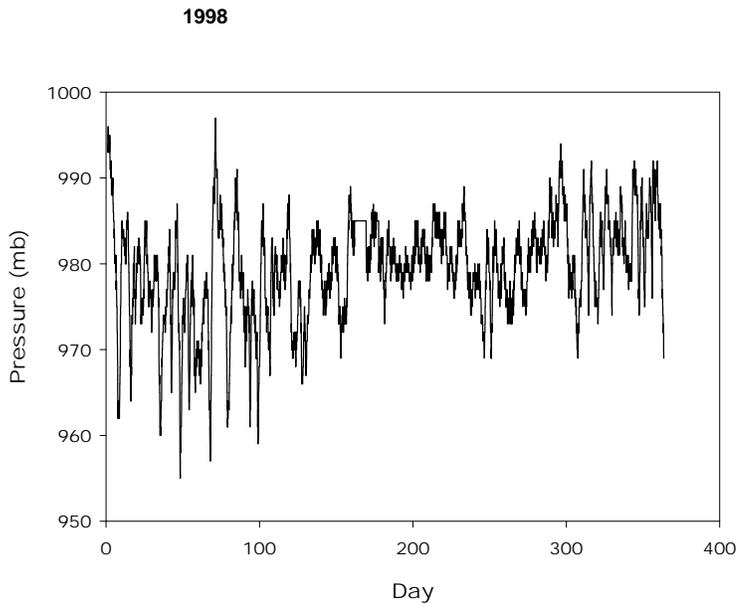


Figure 14 Pressure time series near Oak Ridge, TN. The elevation is 385 m, so the mean annual pressure is 985 mb, rather than 1013 mb, the mean sea level value.

Variation of density with Height

Manipulation of gas laws also allows us to explore how density varies with height. We will find this to be important with regards to differences in momentum transfer at low and high elevations (ρu):

$$\frac{1}{\rho} \frac{d\rho}{dz} = -\frac{1}{T} \left(\frac{dT}{dz} + \frac{g}{R} \right)$$

$$\frac{1}{\rho} \frac{d\rho}{dz} = -\frac{g}{RT_m}$$

$$\rho = \rho(0) \exp\left(-\frac{gz}{RT_m}\right)$$

Standard atmospheric pressure at sea-level is 1013 mb or 101325 Pa

The Universal Gas Constant, R, 8.31441 J K⁻¹ mol⁻¹

CO₂ density

$$\rho_c = [CO_2] \cdot 10^{-6} \rho_a \frac{M_c}{M_a} = [CO_2] \cdot 10^{-6} \frac{P \cdot M_c}{R \cdot T_k}$$

ozone in ppb

$$\rho_{o_3} = [O_3] \cdot 10^{-9} \rho_a \frac{M_{o_3}}{M_a} = [O_3] \cdot 10^{-9} \frac{P \cdot M_{o_3}}{R \cdot T_k}$$

Application of Class Material

You' ve been assigned the job of measuring horizontal or vertical humidity gradients.

1. You have a relative humidity sensor with a 1% accuracy. How well can you measure vapor pressure at temperatures of 5, 15, 25 and 35 C, and relative humidities of 25, 50 and 75%?

2. You have a wet bulb psychrometer with a 0.5 C accuracy. How well can you measure vapor pressure at wet bulb temperatures of 5, 15, 25 and 35 C?

Bibliography:

Campbell, G.S. and Norman, J.M. 1998. An Introduction to Environmental Biophysics. Springer.

Hayden, B.P. 1998. Ecosystem feedbacks on climate at the landscape scale. Philosophical Transactions of the Royal Society of London, B. 353, 5-18.

Jones, H.G. 1992. Plants and Microclimate. Cambridge Press

Biometeorology, ESPM 129, Water, humidity and trace gases

Monteith, J.L. and M.H. Unsworth. 1990. Principles of Environmental Physics. E.A. Arnold.

Oke, T.R. 1987. Boundary Layer Climates. Methuen.

Tindall, J.A. and J.R. Kunkel. 1999. Unsaturated Zone Hydrology for Scientists and Engineers. Prentice Hall.624 pp.

EndNote References

Baldocchi DD, Rao KS (1995) Intra-field variability of scalar flux densities across a transition between a desert and an irrigated potato Advection field. *Boundary Layer Meteorology*. **76.**, 109-136.

Schwartz MD, Karl TR (1990) Spring phenology-natures experiment to detect the effect of green-up on surface maximum temperatures. *Monthly Weather Review* **118**, 883-890.