Lecture Ch. 6a

100%. For simplicity, we assume here that clouds form in the atmosphere when the water vapor reaches its saturation value and $\mathcal{H}$= 100%.

- Saturation of moist air
- Relationship between humidity and dewpoint
  - Clausius-Clapeyron equation
- Dewpoint
  - Temperature
  - Depression
- Isobaric cooling

Curry and Webster, Ch. 6
For Wednesday: Homework, Ch. 6, Prob. 4, 6; Read Ch. 7

How does saturation occur?

- By increasing water vapor
  - Evaporation of water at surface
  - Evaporation of falling rain
- By cooling
  - Isobaric
  - Radiative cooling of rising air
- By mixing of two unsaturated air parcels

Curry and Webster, Ch. 6

Saturation of Moist Air

- Dew point temperature

The temperature at which saturation is reached in an isobaric cooling process is the *dew-point temperature*, which is illustrated in Figure 6.1a. The dew-point temperature, denoted by T_D, can be defined by

$$e = e_s(T_D) \qquad (6.14)$$

or equivalently by

$$w_v = w_s(T_D) \qquad (6.15)$$

Curry and Webster, Ch. 6

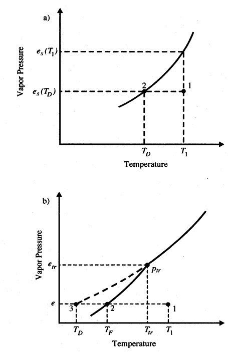


Figure 6.1 a) Relationship between temperature and vapor pressure in an isobaric cooling process. Air initially at temperature T_1 (point 1) is cooled isobarically until it reaches saturation (point 2). The temperature at point 2 defines the dew-point temperature, T_D. b) Air at T_1 (point 1) cools isobarically until it reaches saturation. If the saturation is reached with respect to ice (point 2), the temperature is called the frost point, T_F.

Saturation of Moist Air

- Clausius-Clapeyron equation at dew point

$$\frac{dp}{dT} \approx \frac{L_{lv}}{T v_v} \qquad (4.18)$$

$$v_v = \frac{R_v T}{p}$$

$$\frac{dp}{dT} = \frac{L_{lv} p}{R_v T^2} \qquad (4.19)$$

$$\frac{dp}{p} = \frac{L_{lv}}{R_v T^2} dT$$

$$d \ln p = \frac{L_{lv}}{R_v T^2} dT$$

$$\frac{d \ln p}{dT} = \frac{L_{lv}}{R_v T^2}$$

$$\frac{d(\ln e)}{dT_D} = \frac{L_{lv}}{R_v T_D^2} \qquad (6.18)$$

Clausius Clapeyron

- Recall by integration between two temperatures we had

$$\int_{e_1}^{e_2} d(\ln e) = \int_{T_1}^{T_2} \frac{L_{lv}}{R_v T^2} dT \qquad (4.21)$$

to yield

$$\ln \frac{e_2}{e_1} = -\frac{L_{lv}}{R_v}\left(\frac{1}{T_2} - \frac{1}{T_1}\right) \qquad (4.22)$$

or

$$e_2 = e_1 \exp\left[-\frac{L_{lv}}{R_v}\left(\frac{1}{T_2} - \frac{1}{T_1}\right)\right] \qquad (4.23)$$

Dewpoint and Humidity

- Integrating from ambient to saturation

$$\ln\frac{e_s}{e} = -\ln\mathcal{H} = \frac{L_{lv}}{R_v}\left(\frac{1}{T_D} - \frac{1}{T}\right)$$

or equivalently

$$\mathcal{H} = \exp\left[-\frac{L_{lv}}{R_v}\left(\frac{T-T_D}{T\,T_D}\right)\right] \qquad (6.19)$$

- Dew point depression ($T-T_D$)

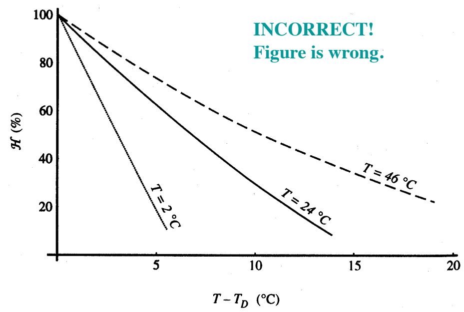
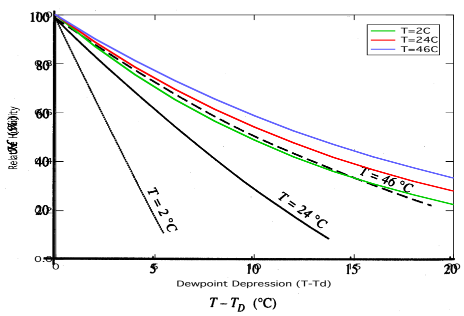


Figure 6.2 Dew–point depression. As the relative humidity increases, the difference between the ambient temperature and the dew-point temperature (i.e., the *dew–point depression*) decreases. As the ambient temperature decreases, the dew–point depression becomes less sensitive to changes in the relative humidity.



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Equivalent Potential Temperature

- Accounts for liquid water heating

$$\theta_e = \theta \exp\left(\frac{L_{lv} w_s}{c_{pd} T}\right) \qquad (6.48)$$

Temperature Metrics

- **Virtual Temperature:** The temperature air would have at the given pressure and density if there were no water vapor in it
- **Potential Temperature:** The temperature a parcel would have if it were brought adiabatically and reversibly to p_0 (usually 1 atm)
- **Virtual Potential Temperature:** The temperature a parcel would have if there were no water vapor in it (only condensed water) and if it were brought adiabatically and reversibly to p_0 (usually 1 atm)
- **Equivalent Temperature:** The temperature that an air parcel would have if all of the water vapor were to condense in an adiabatic isobaric process
- **Equivalent Potential Temperature:** The temperature a parcel would have if all of the water were condensed in an adiabatic isobaric process and if it were brought adiabatically and reversibly to p_0 (usually 1 atm)

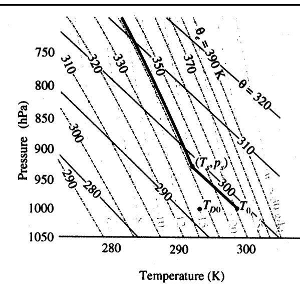

Figure 6.8 Adiabatic ascent of a parcel from p_0. The parcel initially ascends dry adiabatically along the constant potential temperature line that passes through (T_0, 1000 hPa). As the parcel ascends, the saturation mixing ratio decreases while the actual mixing ratio remains the same. At the point at which the actual mixing ratio of the parcel is equal to the saturation mixing ratio, the parcel becomes saturated. Further lifting of the parcel occurs along the saturated adiabat that passes through the point, (T_s, p_s).

$$\theta_e = T_e\left(\frac{p_0}{p}\right)^{R_d/c_p}$$

Cumulus Cloud Base Altitude Calculator

Cloud Base Altitude = ((((temperature - dew point) / 4.5) * 1000) + measure station altitude)

Assumes:
-The rate at which air cools as it rises is averaged at 5.5°F per 1000 feet
-The dew point also decreases at about 1.0°F over the same distance.

http://www.csgnetwork.com/estcloudbasecalc.html