Lecture Ch. 6

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
- Moist adiabatic ascent of air
- Equivalent temperature
- Aerological diagrams

Curry and Webster, Ch. 6

Condensed (Liquid) Water

- **Ch. 4: Will H_2O condense?**
  - Clausius-Clapeyron describes $e_{sat}(T)$
- **Ch. 5: What will H_2O condense on?**
  - Kohler describes droplet formation
- **Ch. 6: How much H_2O will condense?**
  - Dew point temperature provides metric
- **Ch. 7: How does condensed H_2O affect stability?**
  - Conditional stability is affected by moist adiabats
- **Ch. 8: What happens to condensed H_2O?**
  - Precipitation processes

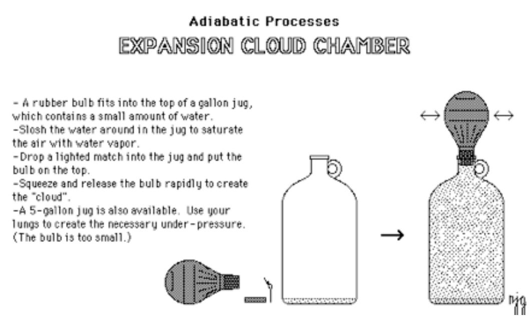
How does saturation occur?

- 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

http://www.youtube.com/watch?v=XH_M4jItiKw

Curry and Webster, Ch. 6

Cloud in a Jar Demonstration



http://groups.physics.umn.edu/demo/old_page/demo_gifs/4B70_20.GIF

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

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

or equivalently by

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

Curry and Webster, Ch. 6

Saturation of Moist Air

- Clausius-Clapeyron equation at dew point

$$\frac{dp}{dT} \approx \frac{L_{lv}}{T\upsilon_v} \quad (4.18)$$

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

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

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

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

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

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

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)

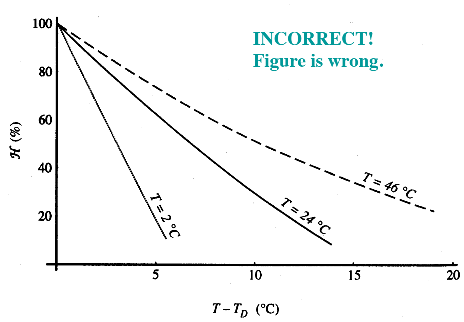


Figure 6.2 Dew–point depression. As the relative humidity increases, the difference between the ambient temperature and the dew-point temperature (i.e., the *dew–point depression*) decreases. As the ambient temperature decreases, the dew–point depression becomes less sensitive to changes in the relative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

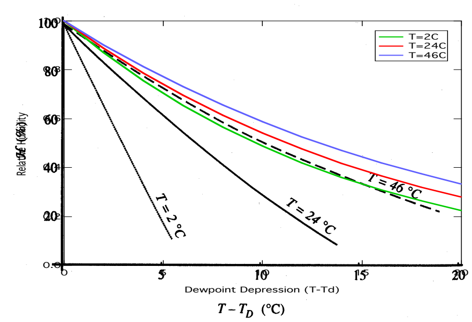


Figure 6.2 Dew–point depression. As the relative humidity increases, the difference between the ambient temperature and the dew-point temperature (i.e., the *dew–point depression*) decreases. As the ambient temperature decreases, the dew–point depression becomes less sensitive to changes in the relative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

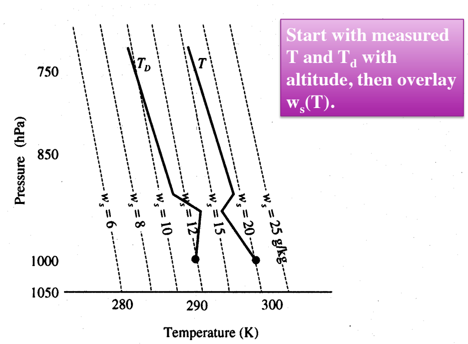


Figure 6.7 Determination of w, w_s, and T_D given the vertical profiles of temperature and dew-point temperature.

Equivalent Potential Temperature

The simplest possible case is that in which saturation conditions are maintained, ice is not present, and heat capacity of the water vapor and condensed water are neglected relative to that of dry air. Using these approximations, the entropy equation (6.13) becomes:

$$0 = c_{pd}\, d(\ln T) - R_d\, d(\ln p) + d\left(\frac{L_{lv} w_s}{T}\right) \qquad (6.47)$$

for a dry adiabatic process from (2.63)

$$c_{pd}\, d(\ln\theta) = c_{pd}\, d(\ln T) - R_d\, d(\ln p)$$

$$-d\left(\frac{L_{lv} w_s}{T}\right) = c_{pd}\, d(\ln\theta)$$

This expression is integrated to a height in the atmosphere where all of the water vapor has been condensed out by adiabatic cooling. The corresponding temperature is called the ***equivalent potential temperature*, θ_e**. Integration of

$$-L_{lv}\int_{w_s}^{0} d\left(\frac{w_s}{T}\right) = c_p \int_{\theta}^{\theta_e} d(\ln\theta)$$

Equivalent Potential Temperature

- Accounts for liquid water heating

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

Although approximate, (6.48) retains the condensation of water vapor provides energy to the moist air and increases its temperature relative to what the temperature would have been in dry adiabatic ascent.

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)

T vs. ln(p)

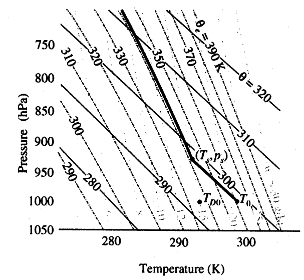
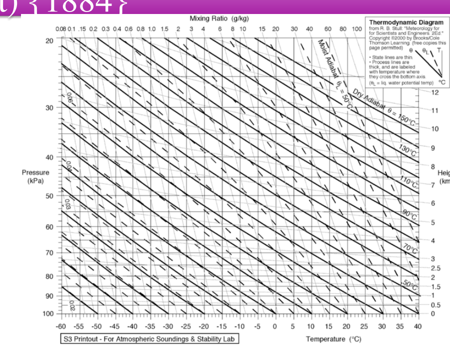


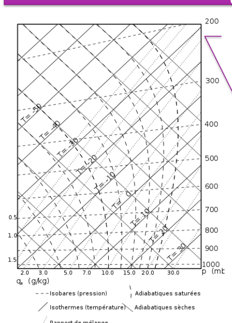
Figure 6.8 Adiabatic ascent of a parcel from p_0. The parcel initially ascends dry adiabatically along the constant potential temperature line that passes through (T_0, 1000 hPa). As the parcel ascends, the saturation mixing ratio decreases while the actual mixing ratio remains the same. At the point 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).

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

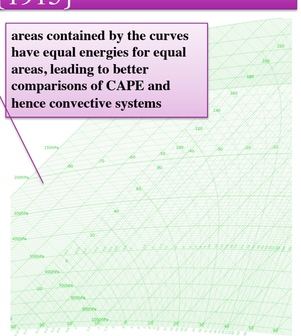
Emagram (T is vertical; thetas slant left) {1884}



Tephigram (T slants right; thetas slants left) {1915}



areas contained by the curves have equal energies for equal areas, leading to better comparisons of CAPE and hence convective systems



Stuve diagram (T is vertical) {1927}

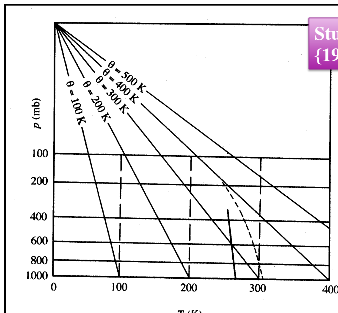
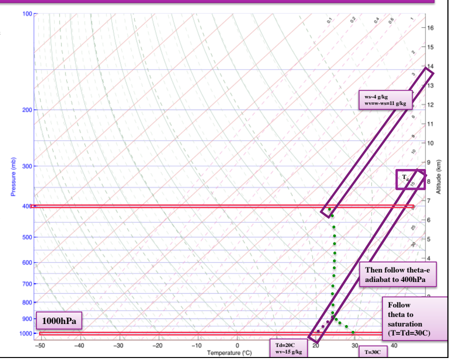


Figure 6.6 Construction of the pseudo-adiabatic chart.

the surface. Thus the ordinate may be proportional to $-\ln p$ (the *Emagram*) or to p^{R/c_p} (the *Stuve diagram*). The Emagram has the advantage over the Stuve diagram in that area on the diagram is proportional to energy. Before the advent of computers,

Skew-T Log P (T slants right) {1947}

- All temperatures are equal at the horizontal 1000 mb level near the bottom.
  - red lines are isotherms,
  - dashed green lines are the equivalent potential temperature;
  - solid green lines are the potential temperature
- Blue lines are the isobars (scale on the left side)
- Dashed purple lines are the **saturated** humidity mixing ratio ($w_s(T)$, i.e. nearly parallel to T)



How to find dew point and frost point?
Cool *isobarically* until reaching saturation.

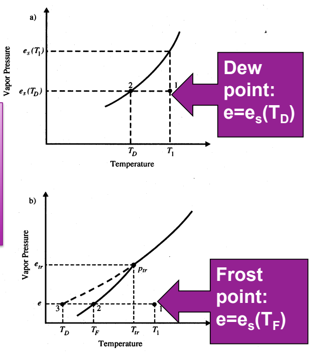


Figure 6.1 a) Relationship between temperature and vapor pressure in an isobaric cooling process. Air initially at temperature T_1 (point 1) is cooled isobarically until it reaches saturation (point 2). The temperature at point 2 defines the dew-point temperature, T_D. b) Air at T_1 (point 1) cools isobarically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

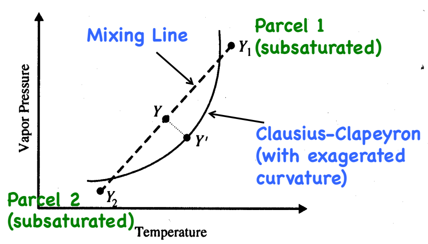


Figure 6.3 Adiabatic isobaric mixing and condensation. Two air masses with (e, T) given by points Y_1 and Y_2 mix, resulting in a single air mass with (e, T) given by point Y. Since $\mathcal{H} > 1$ at this point, water will condense, and the temperature of the air mass will increase while the vapor pressure decreases. Condensation will continue until the temperature and vapor pressure of the air mass coincide with the saturation vapor pressure curve (point Y').

N.B. For mixture to be supersaturated, need two parcels very close to saturation at two somewhat different temperatures.

Enthalpy Change for Phase+Temperature Change

The exact differential of the enthalpy, dH, where $H = H(T, p, m_d, m_v, m_l)$, can be expanded as follows:

$$dH = \left(\frac{\partial H}{\partial T}\right) dT + \left(\frac{\partial H}{\partial p}\right) dp + \left(\frac{\partial H}{\partial m_d}\right) dm_d + \left(\frac{\partial H}{\partial m_v}\right) dm_v + \left(\frac{\partial H}{\partial m_l}\right) dm_l$$

If the system is closed, then $dm_d = 0$ and $dm_v = -dm_l$, and therefore

$$dH = \left(\frac{\partial H}{\partial T}\right) dT + \left(\frac{\partial H}{\partial p}\right) dp + \left[\left(\frac{\partial H}{\partial m_v}\right) - \left(\frac{\partial H}{\partial m_l}\right)\right] dm_v \qquad (6.1a)$$

Since $(h_v - h_l) = L_{lv}$ (Section 4.3), we have

$$dH = \left(\frac{\partial H}{\partial T}\right) dT + \left(\frac{\partial H}{\partial p}\right) dp + L_{lv}\, dm_v \qquad (6.1b)$$

Enthalpy Change for Phase+Temperature Change

not distinguish between them. In Section 2.3, we found that $\partial H/\partial p = 0$ for an ideal gas. For liquid water, $\partial H/\partial p \neq 0$, but the value is small and thus neglected here. We can therefore write (6.1) as

$$dH = \left(m_d c_{pd} + m_v c_{pv} + m_l c_l\right) dT + L_{lv}\, dm_v \qquad (6.2b)$$

In the atmosphere, the mass of water vapor is only a few percent of the mass of dry air (Section 1.1), and the mass of condensed water is a small fraction of the mass of water vapor. Thus $m_d >> m_v >> m_l$ and we can approximate (6.2b) by

$$dH \approx m_d c_{pd}\, dT + L_{lv}\, dm_v \qquad (6.3)$$

The enthalpy of a system consisting of moist air and a liquid water cloud is not only a function of temperature (as was the ideal gas), but also a function of the latent heat associated with the phase change. In intensive form, we have

$$dh \approx c_{pd}\, dT + L_{lv}\, dw_v \qquad (6.4)$$

Wet-Bulb Temperature

Consider a system composed of unsaturated moist air plus rain falling through the air. Because the air is unsaturated, the rain will evaporate. If there are no external heat sources ($\Delta q = 0$), and the evaporation occurs isobarically ($dp = 0$), we can write an adiabatic, isobaric (or *isenthalpic*) form of the enthalpy equation (6.20) as

$$0 = dh = c_p\, dT - L_{lv}\, dw_l = c_p\, dT + L_{lv}\, dw_s \qquad (6.23)$$

where c_p can be approximated as the dry–air value, or alternatively the contributions from water vapor and liquid water can be incorporated following (6.2a). Since

If we allow just enough liquid water from the rain to evaporate so that the air becomes saturated, we can integrate (6.23)

$$c_p \int_T^{T_W} dT = -L_{lv} \int_{w_l}^{0} dw_s$$

$$c_p\left(T_W - T\right) = -L_{lv}\left[w_s\left(T_W\right) - w_v\right]$$

where w_l represents the amount of water that must be evaporated to bring the air to saturation. During the evaporation process, latent heat is drawn from the atmosphere, and the final temperature, referred to as the *wet-bulb temperature* (T_W), is cooler than

$$T_W = T - \frac{L_{lv}}{c_p}\left[w_s\left(T_W\right) - w_v\right] \qquad (6.24)$$

Earth's Annual Global Mean Energy Budget

J. T. Kiehl and Kevin E. Trenberth
National Center for Atmospheric Research,* Boulder, Colorado

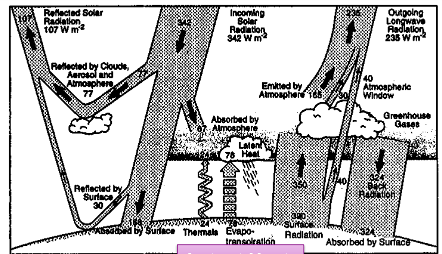
ABSTRACT

The purpose of this paper is to put forward a new estimate, in the context of previous assessments, of the annual global mean energy budget. A description is provided of the source of each component to this budget. The top-of-atmosphere shortwave and longwave flux of energy is constrained by satellite observations. Partitioning of the radiative energy throughout the atmosphere is achieved through the use of detailed radiation models for both the longwave and shortwave spectral regions. Spectral features of shortwave and longwave fluxes at both the top and surface of the earth's system are presented. The longwave radiative forcing of the climate system for both clear (125 W m^{-2}) and cloudy (155 W m^{-2}) conditions are discussed. The authors find that for the clear sky case the contribution due to water vapor to the total longwave radiative forcing is 75 W m^{-2}, while for carbon dioxide it is 32 W m^{-2}. Clouds alter these values, and the effects of clouds on both the longwave and shortwave budget are addressed. In particular, the shielding effect by clouds on absorption and emission by water vapor is as large as the direct cloud forcing. Because the net surface heat budget must balance, the radiative fluxes constrain the sum of the sensible and latent heat fluxes, which can also be estimated independently.

In final form 5 August 1996.

Bulletin of the American Meteorological Society

Radiation Transport in the Atmosphere



Latent Heat accounts for 78 W/m^2!

Figure 12.2 Estimated annual mean [...] for the Earth. Units are W m^{-2} (Kiehl and Trenberth, 1997).

What did we learn in Ch. 6?

- Cloud formation – what causes saturation?
  - Cooling to decrease $e_s(T)$
  - Increasing water vapor to increase e
  - Mixing of two nearly-saturated parcels
- Enthalpy balance: **How <u>much</u> H_2O will condense?**
  - Latent heat from evaporation causes cooling
  - Latent heat from condensation causes warming
- Dew point (T_D) and frost point (T_F) temperatures
  - Cool isobarically until saturation is reached
  - RH scales with dewpoint depression (T-T_D)
- L_{lv} affects energy balance (Kiehl+Trenberth, 1997)