Lecture Ch. 7a

- Stability
- CAPE
- Review of Ch.7 Concepts
  - "Homework" Ch. 7, Prob. 3 for discussion
- Cloud Classification
- Precipitation Processes

Curry and Webster, Ch. 7, 8
For Monday: Read Ch. 8

Dry/Moist/Saturated

- Dry (RH=0%)
  - In practice, 0%<RH<100% (moist air) can sometimes be approximated as "dry"
- "Moist" (0%<RH<100%)
  - Example: saturated air with some dry air entrained into it (7.27)
- Saturated (RH≥100%)
  - Some liquid water is present
  - Approximate using "equivalent" or "liquid water" potential temperature

Lapse Rate and Stability

- Lapse Rate (Γ): helps to define the stability of the atmosphere.
- Degree of stability has consequences for atmospheric mixing, e.g. dispersion of pollutants.
- Stability:
  - Superadiabatic (unstable)
    $\Gamma_{env} > \Gamma_{ad}$
  - Subadiabatic (stable)
    $\Gamma_{env} < \Gamma_{ad}$
  - Neutral
    $\Gamma_{env} = \Gamma_{ad}$

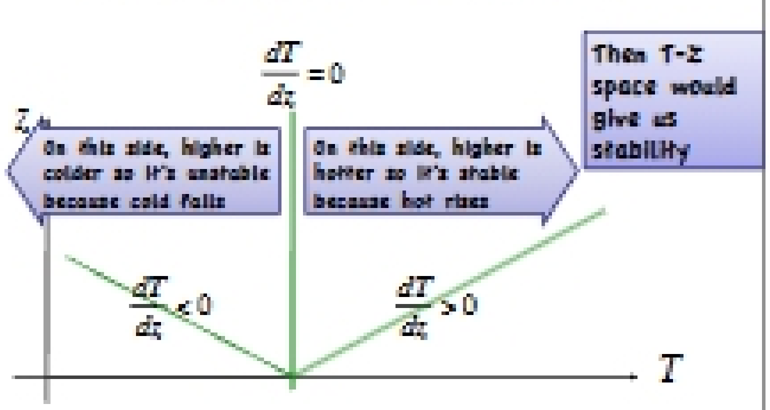
Why do we need stability?

- Buoyancy tells us *that* air rises
- Lapse rates tell us *how* air rises
- Stability tells us *whether* rise will continue or reverse

Static stability:
"If a parcel of air is displaced vertically, does it return to its original level (stable) or does it continue to move away from it (unstable)"

Suppose!

If atmosphere had P~constant with z, then we would find T~constant with z



BUT WAIT!

Pressure is NOT constant with z
So T is not constant with z

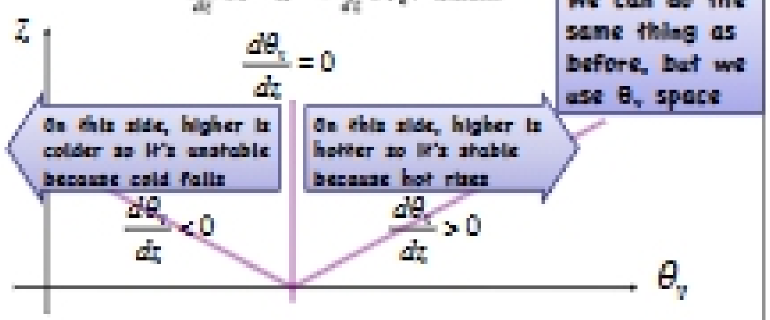
But what is constant with z?

Unsaturated Stability Criteria

$$\frac{d\theta_v}{dz} > 0 \quad \text{or} \quad -\frac{dT_v}{dz} < \Gamma_d : \text{ stable}$$

$$\frac{d\theta_v}{dz} = 0 \quad \text{or} \quad -\frac{dT_v}{dz} = \Gamma_d : \text{ neutral} \qquad (7.20)$$

$$\frac{d\theta_v}{dz} < 0 \quad \text{or} \quad -\frac{dT_v}{dz} > \Gamma_d : \text{ unstable}$$



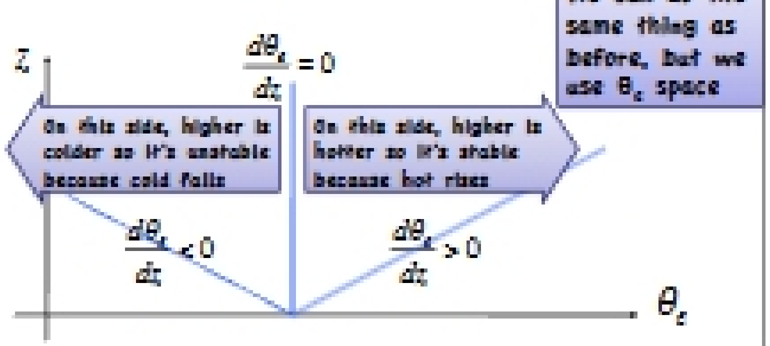
BUT WAIT!

θ_v is NOT constant with z if atmosphere is saturated.

But what is constant with z?

Saturated Stability Criteria

i) the saturated layer will be stable if $d\theta_e/dz > 0$;
ii) the saturated layer will be neutral if $d\theta_e/dz = 0$;
iii) the saturated layer will be unstable if $d\theta_e/dz < 0$.



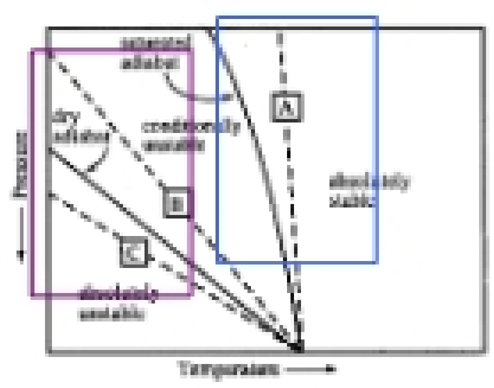


Figure 7.1 Regions of stability, instability, and conditional stability illustrated on an aerological diagram. When the environmental lapse rate is less than the saturated adiabatic lapse rate (e.g., lapse rate A), the atmosphere is absolutely stable. When the environmental lapse rate is greater than the saturated lapse rate, but less than the dry adiabatic lapse rate (e.g., lapse rate B), the atmosphere is conditionally stable. When the environmental lapse rate is greater than the dry adiabatic lapse rate (e.g., lapse rate C), the atmosphere is absolutely unstable.

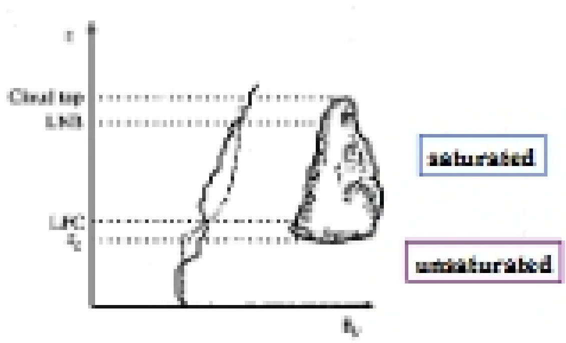


Figure 8.21 Typical temperature profiles in a convective environment. The solid profile represents the environmental temperature; the dashed profile corresponds to the temperature within the cloud. The cloud base forms near the lifting condensation level, z_c. Near the cloud base, the temperature increases more rapidly with height in the cloud than in the surroundings, resulting in a relatively large temperature difference between the environmental temperature and the cloud interior temperature. A cloud that reaches the level of free convection (LFC) will accelerate upwards until it reaches the level of neutral buoyancy (LNB), where the environmental temperature is equal to the interior cloud temperature.

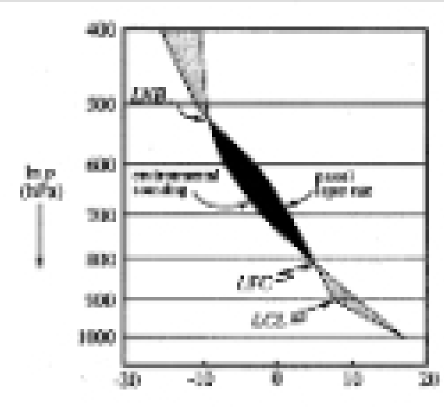


Figure 7.2 Convective instability illustrated on an aerological diagram. The dashed line represents the environment (T) and the solid line represents the parcel (T'). Below 810 mb and above 530 mb, energy is required to lift the parcel. Above 810 mb and below 530 mb, the parcel accelerates freely. The dark shaded area represents the convective available potential energy (CAPE), while the two lightly shaded areas represent the convection inhibition energy (CINE).

CAPE

The amount of energy available for the upward acceleration of a particular parcel is called the convective available potential energy (CAPE). On a thermodynamic diagram whose area is proportional to energy (e.g., the tephigram; see Section 6.8), CAPE is proportional to the area enclosed by the two curves that delineate the temperatures of a parcel and its environment, as illustrated by the darker shaded region in Figure 7.2. The amount of CAPE of a parcel lifted from a height z (at or above the LFC) to the LNB is given by the vertical integral of the buoyancy force between these levels

$$CAPE(z) = \int_{z}^{LNB} g\frac{\rho - \rho'}{\rho'}\,dz \qquad (7.24)$$

where the units of CAPE are J kg^{-1}. If the environment is in hydrostatic equilibrium we can use (7.20) and (1.33) to obtain

$$CAPE(p) = \int_{p(LNB)}^{p} R_d\left(T_v' - T_v\right) d(\ln p) \qquad (7.25)$$

CAPE is defined only for parcels that are positively buoyant somewhere in the vertical profile. The term convective inhibition energy (CINE) is analogous to CAPE but refers to a negative area on the thermodynamic diagram.

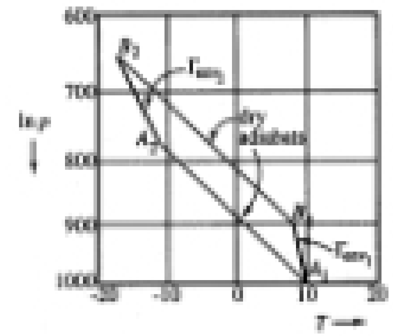


Figure 7.3 An initially stable layer A_1B_1 is made less stable as a result of dry adiabatic ascent.

Figure 7.5 Vertical mixing of air parcels, m_1 and m_2, without condensation. Two air parcels, initially at different pressure levels, mix at an intermediate pressure level. The potential temperature of the mixture is a mass-weighted average of the individual parcels' potential temperatures. Mixing of an entire layer results in a constant potential temperature $\bar{\theta}$ throughout the layer. This destabilizes an initially stable layer and stabilizes an initially unstable layer. Because the dry adiabat corresponding to $\bar{\theta}$ does not intersect the average mixing ratio line, $\bar{w}_v$, the mixing process is dry adiabatic and no condensation occurs.

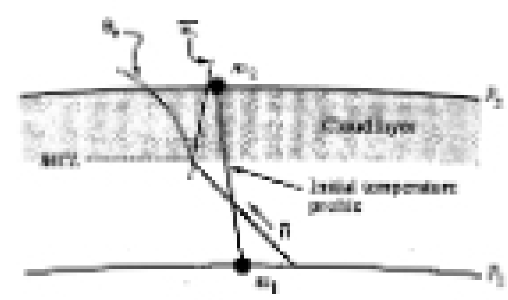


Figure 7.6 Vertical mixing of air parcels, m_1 and m_2, with condensation. If the mixing of two air parcels results in an average potential temperature, $\bar{\theta}$, that intersects the average mixing ratio line, $\bar{w}_v$, then from the level of intersection upwards, condensation will occur and the final temperature distribution will follow a saturated adiabat, θ_e. The lapse rate below the cloud layer moves towards the dry adiabatic lapse rate, while the lapse rate within the cloud layer moves towards the saturated adiabatic lapse rate.

Chapter 7, Prob. 3

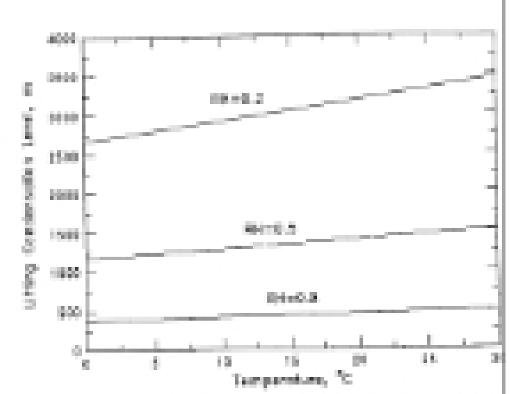
3. Suppose that the environmental lapse rate is dry adiabatic, with a temperature of 280 K at 900 hPa, and a relative humidity of 50%. Consider a parcel of saturated air at 900 hPa at 280 K, initially at rest. If this parcel is given an upward displacement, it will be positively buoyant and will continue to ascend. Neglecting entrainment and aerodynamic resistance, calculate the parcel's upward velocity at 700 hPa, assuming the following:
a) elementary parcel theory without including the virtual temperature correction;
b) elementary parcel theory including the virtual temperature correction;
c) parcel theory with a correction for the weight of condensed water, assuming full adiabatic water content.

$$CAPE(p) = \int_{p(LNB)}^{p(LFC)} R_d\left(T_v' - T_v\right) d(\ln p) \qquad (7.25)$$

- Use a mean moist adiabatic lapse rate of 5.77K/km (valid for p = 800 hPa, T = 276 K; assume constant); dry adiabat 9.77K/km (from http://157.82.240.165/~masaki/comp/calc/index.html)
  - a) use Eq. 7.25 with T to get CAPE= 280.32, assume CAPE=KE (upper bound), v=23.7 m/s.
  - b) use Eq. 7.25 with Tv to get CAPE=331.2, v=25.8 m/s.
  - c) use Fig. 8.2 to get w=1.2g/kg then Eq. 7.25 to get CAPE, so v=20.0 m/s.
- **CAPE is an upper bound on the potential energy that can ever be converted into kinetic energy of a rising buoyant parcel.**

Lifting Condensation Level

- Lifting condensation level varies with initial relative humidity and is a weak function of initial temperature



Seinfeld and Pandis, Fig. 15.11