Radiative transfer and climate III

Reading: GPC Ch3 (omit section 3.11 except for fig 3.20, which we will cover in lecture).

Outline:

- Dry adiabatic lapse rate, potential temperature
- Moist adiabatic lapse rate
- Static stability: dry and moist cases
- Radiative-convective equilibrium
- Flux divergence (convergence) and cooling (heating)

Lapse rate and potential temperature

Molecular mixing (the result of random motions of individual molecules) is important only within a centimeter of the surface and above 100 km. At all the other levels virtually all the vertical mixing is accomplished by the exchange of well defined air parcels with horizontal dimensions ranging from a few cm to the global scale. Let us consider a ***dry air parcel*** of infinitesimal dimensions that has the following properties:

- Thermally insulated - does not gain or lose energy to the environment (***adiabatic transformation***). Most atmospheric processes are fast enough to be nearly adiabatic.
- Same pressure as the environment, which is assumed to be in hydrostatic equilibrium (the mechanical adjustment to equilibrium is much faster than the thermal adjustment).
- Moving slowly enough so that its kinetic energy is negligible.

The internal energy of a unit mass of this parcel remains constant:

c_pT (heat stored) + gz (potential energy at height z) $= \text{const} \Rightarrow -(dT/dz)_{parcel} = g/cp = \Gamma_d$

c_p is the specific heat of the air at const. pressure, 1004 J/(kg K)
g is acceleration due to gravity

Γ_d is the ***dry adiabatic lapse rate*** = it is the change of the temperature of such a parcel as it moves vertically in the earth's atmosphere.
For dry air Γ_d=9.8 degrees C/km

The ***potential temperature*** θ is the temperature that a dry parcel would obtain if it is moved adiabatically from its position (at pressure p and temperature T) to a standard pressure p_o:

$$\theta = T\,(p_o/p)^{R/cp} \qquad \text{(C.9)}$$

R is the gas constant (287 J/K.kg) and c_p is the specific heat of the air at constant pressure, $R/c_p=0.286$.

For the derivation, the basic idea is that the energy associated with compressing (expanding) the air parcel is used to heat (cool) the air parcel.

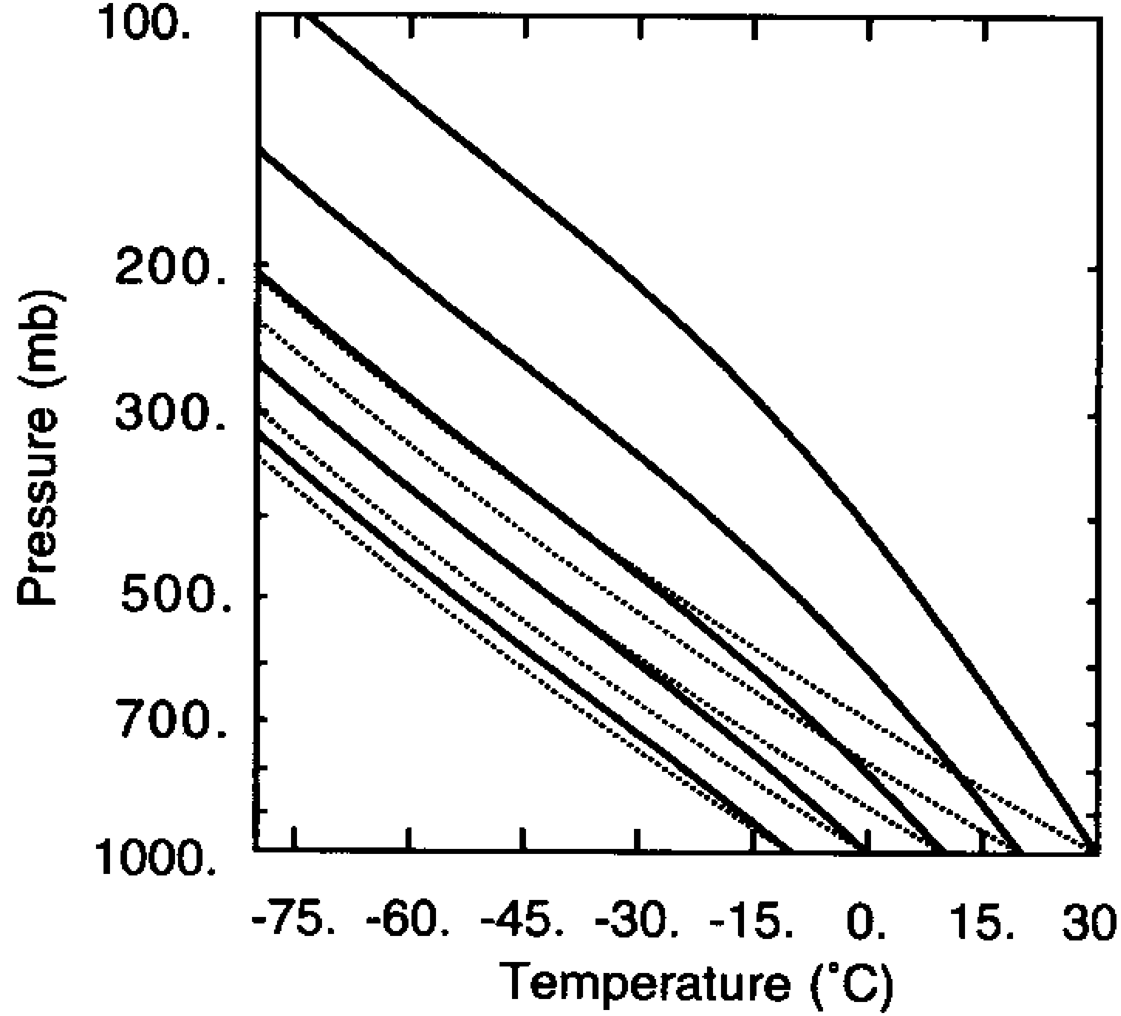
The potential temperature is a *conserved* quantity in adiabatic transformations - i.e. potential temperature of an air parcel remains fixed as long as no heat is added/withdrawn.

For a moist air parcel, as the air parcel is moved up and its temp. decreases, the saturation specific humidity will also decrease. There will come a point where the specific humidity of the parcel exceeds saturation - at this point, the excess water condenses and precipitates out. Heat is released in this process which warms the air parcel.

Latent heat of vaporization (L) is the energy required to vaporize a unit mass of liquid water: L=2.25x10^6 J/kg at 100C, 2.5x10^6 at 0^0C.

Moist (saturated) adiabatic lapse rate Γ_s : the temperature that a *saturated* parcel would attain if it were moved vertically. The energy from condensation is assumed to remain trapped within the air parcel (hence the tranformation is *adiabatic* and *reversible*).

$$\Gamma_s = \frac{\Gamma_d}{1 + \left(L/c_p \right)\left(dq^*/dT \right)}$$



Moist (solid) and dry (dotted) adiabatic lapse rates, for different surface temperatures

Concept of static stability:

for an unsaturated air parcel

Stable situation | Unstable situation

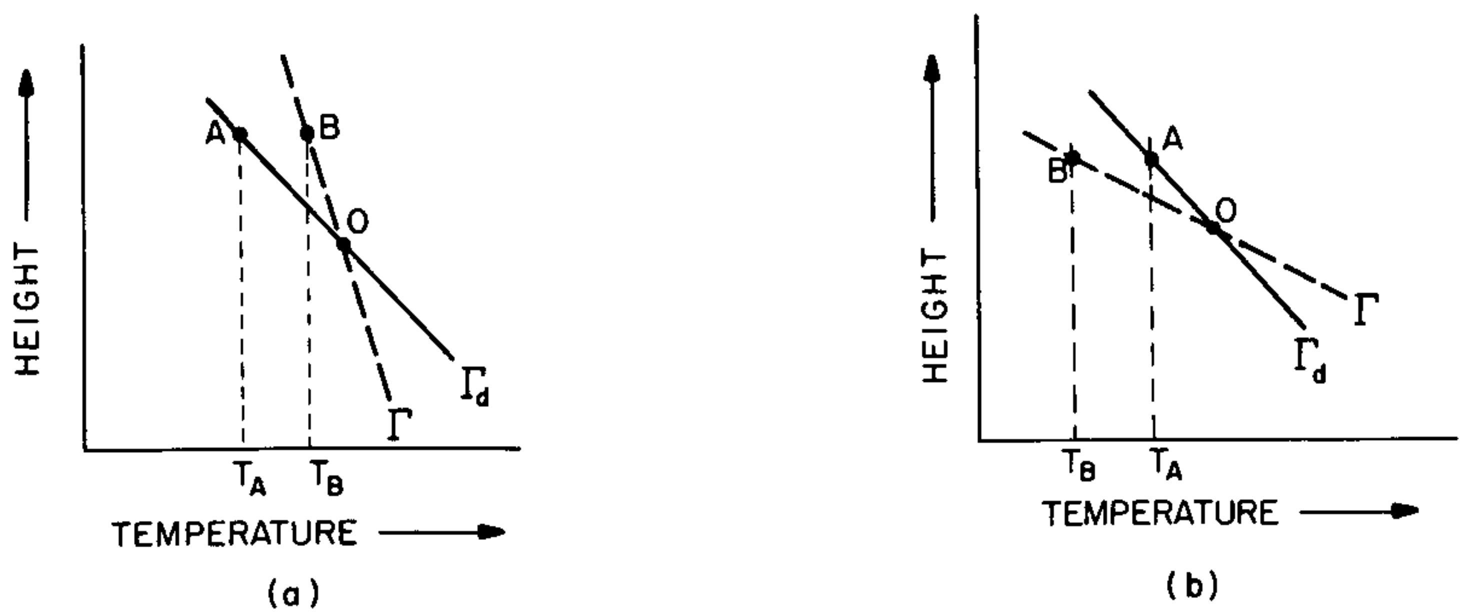
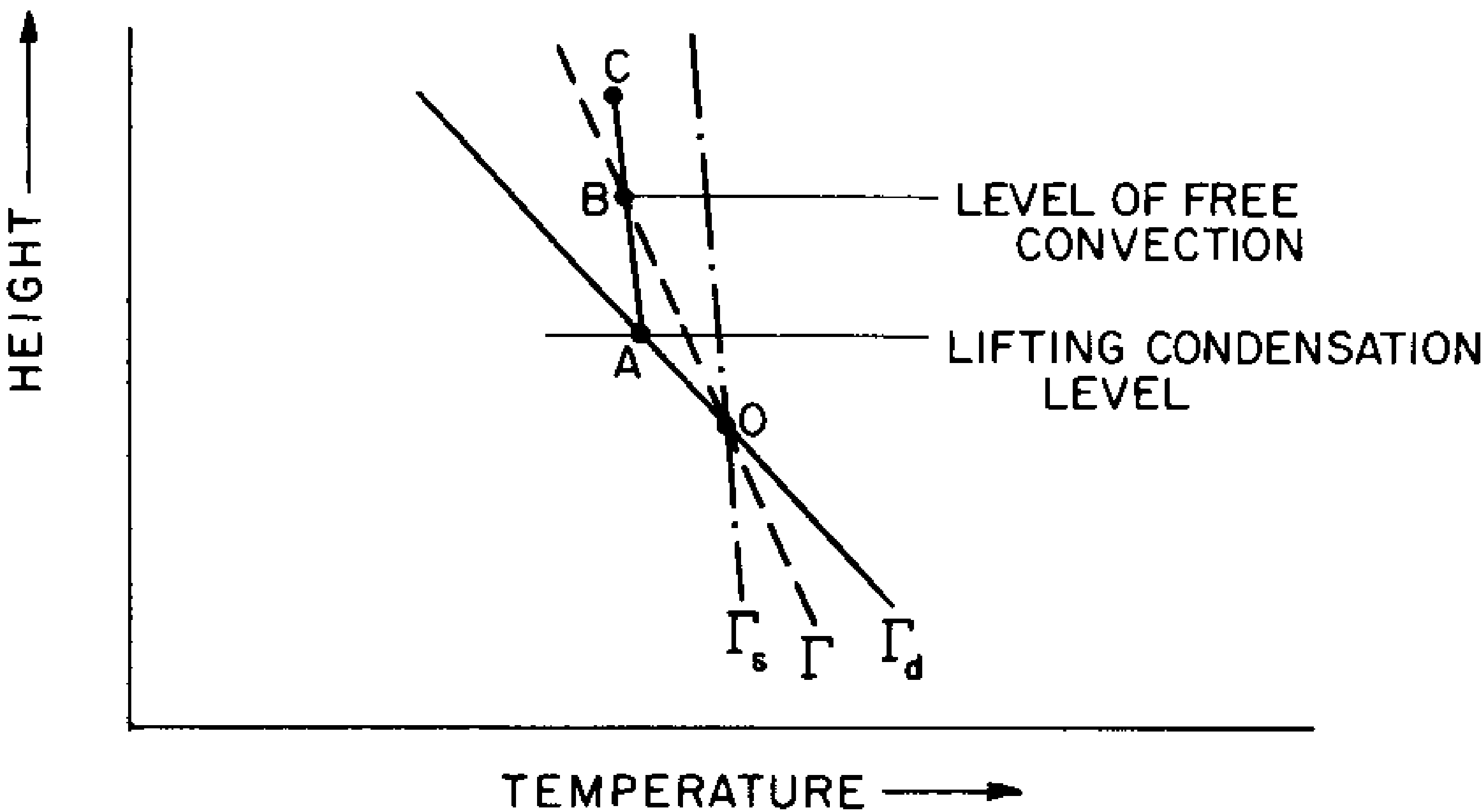
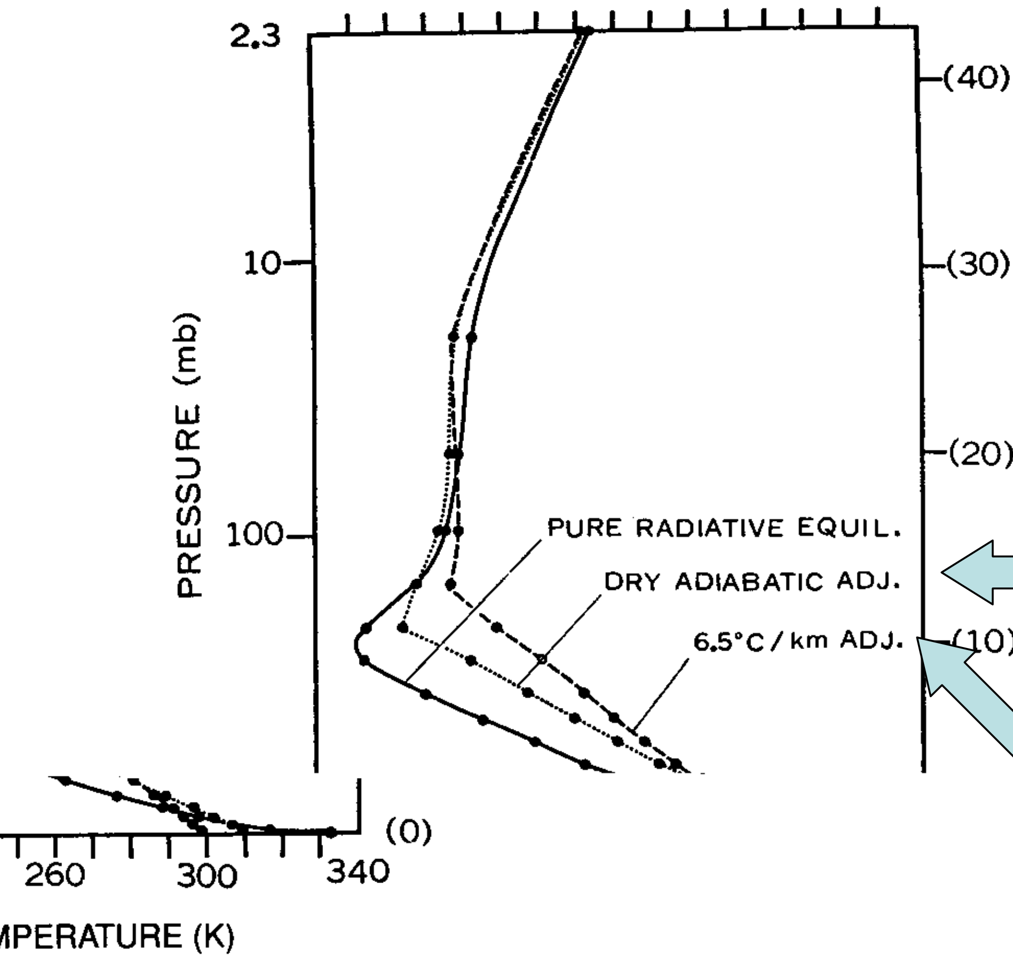


Fig. 2.10 Conditions for (a) positive static stability and (b) negative static stability for unsaturated air. Negative static stability can only exist very close to the ground.

Moist conditional instability



Radiative-***convective*** (or thermal) equilibrium profile:



Troposphere warmer and surface cooler than for radiative equilibrium case

with dry convection

with “moist convection” (and other processes)

for radiative equilibrium, and thermal equilibrium with
om Manabe and Strickler (1964). Reprinted with permis-
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. **3.16** Calculated temperature profiles
ıtes of 9.8°C km^{-1} and 6.5°C km^{-1}. [Fr
om the American Meteorological Society.

The radiative-convective equilibrium model is useful to understand how important are the individual greenhouse gases in determining the vertical profile of temperature.

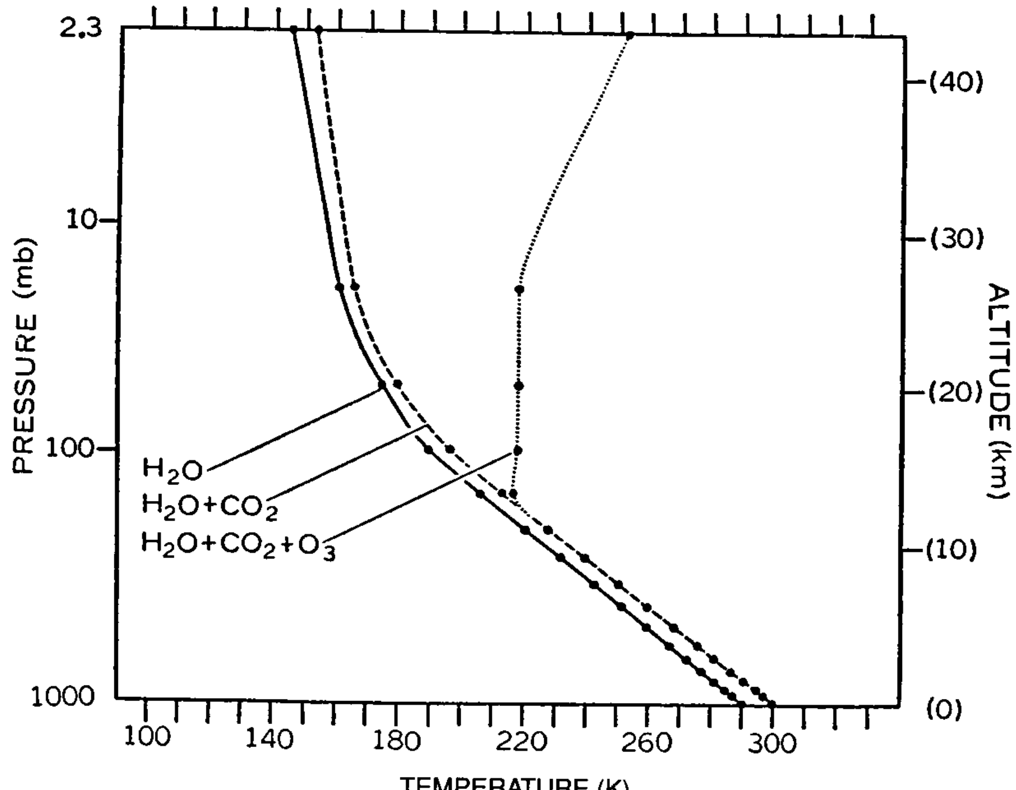
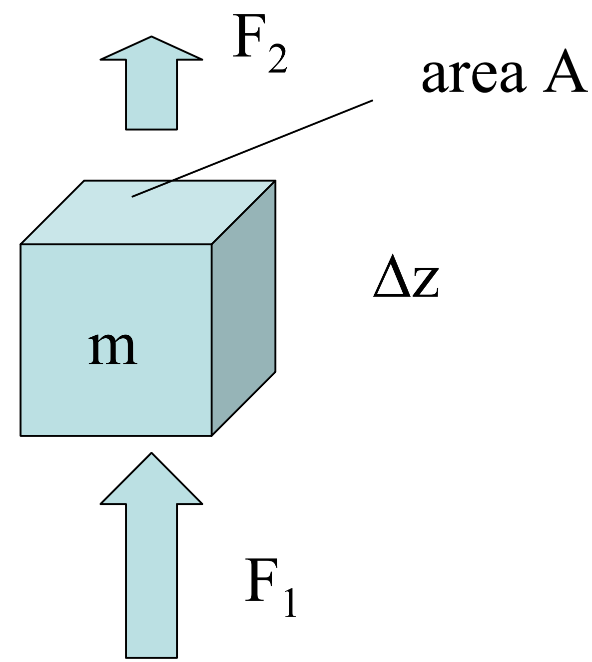


Fig. 3.17 Thermal equilibrium profiles for three cloudless atmospheres obtained with a critical lapse rate of 6.5 K km^{-1}. One atmosphere has water vapor only; one includes water vapor and carbon dioxide; and the third contains water vapor, carbon dioxide, and ozone. [From Manabe and Strickler (1964). Reprinted with permission from the American Meteorological Society.]

Flux divergence (convergence) and heating

Rate of energy input = $-(F_2 - F_1) \times A$

Heating: Mass x c_p x ΔT =
Rate of energy input x Δt



So: $\Delta T / \Delta t = -(1/\rho\, c_p)\, \Delta F / \Delta z$
(at a given level of given pressure)

OR

$$\underbrace{\frac{\partial T}{\partial t}}_{\text{Time rate of change of T}} = -\frac{1}{\rho c_p} \underbrace{\frac{\partial F}{\partial z}}_{\text{“flux divergence”}} \qquad (3.38)$$

Radiative-convective equilibrium model
Heating rates: contributions from H_2O, CO_2, O_3

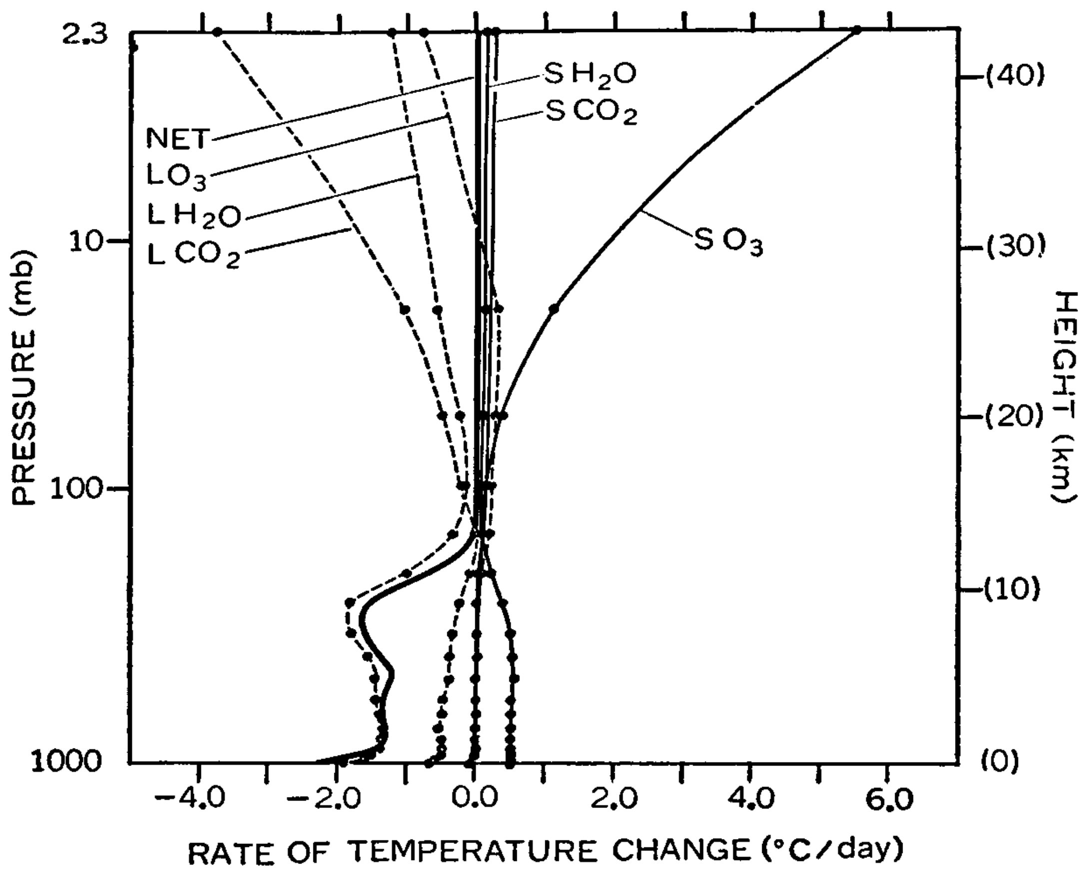


Fig. 3.18 Radiative heating rate profiles for a clear atmosphere. LH_2O, LCO_2, and LO_3 show the ating rates associated with longwave cooling by water vapor, carbon dioxide, and ozone, respectively. e S prefix indicates the heating rate associated with solar absorption by each of these gases. NET is the m of the solar and longwave radiative heating rates contributed by all gases. [From Manabe and Strick- r (1964). Reprinted with permission from the American Meteorological Society.]

$$\frac{\partial T}{\partial t} = -\frac{1}{\rho c_p}\frac{\partial F}{\partial z}$$