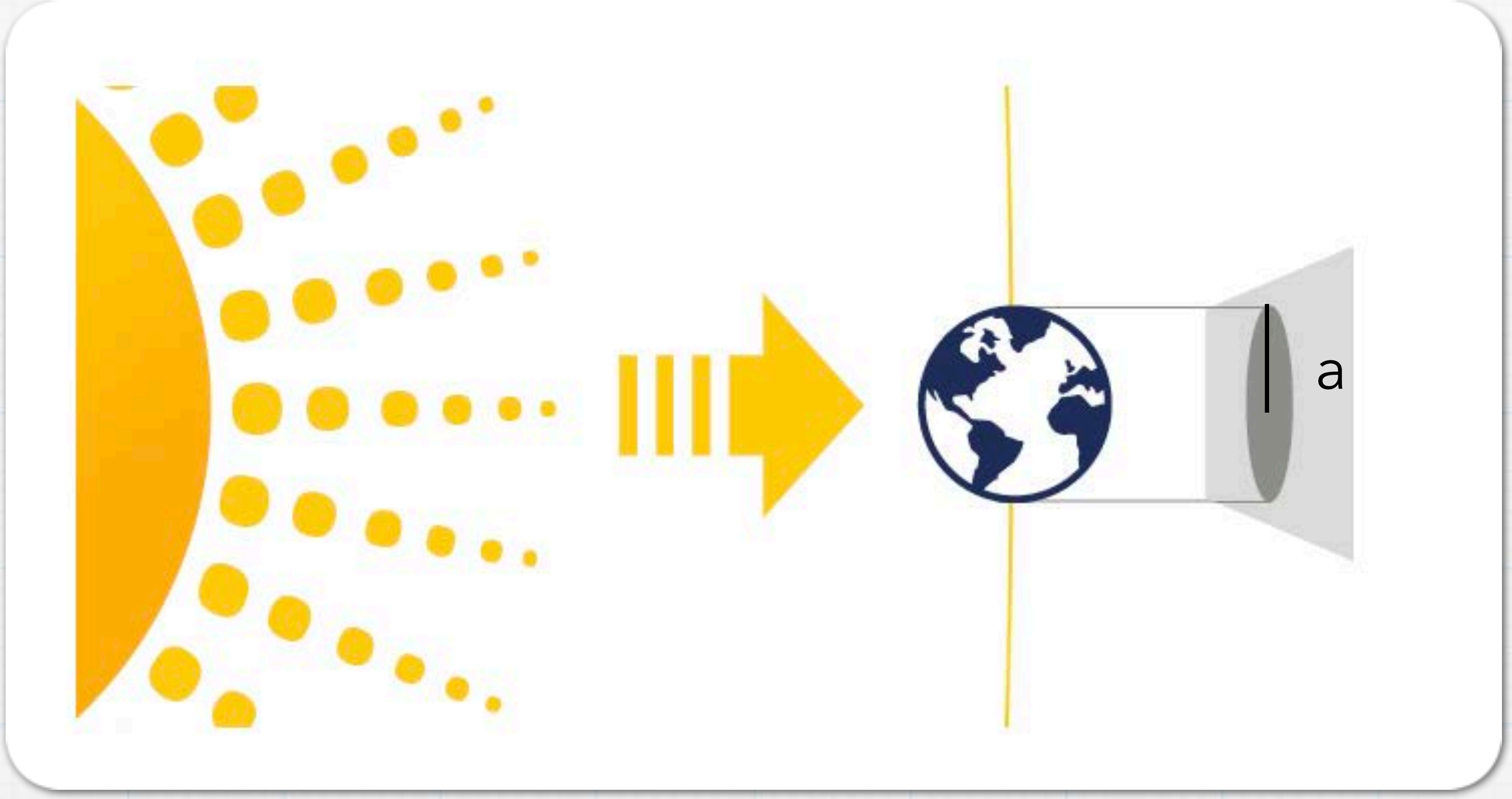
# Atmosphere: #2
# Intro and fundamentals

# 1. Planetary emission temperature

- Solar power incident on the Earth = $S_0 \pi a^2$ = 1.74 x $10^{17}$ W
- So, is this the amount of solar energy that the Earth absorbs?
- NO! A significant fraction is reflected.



# 1. Planetary emission temperature

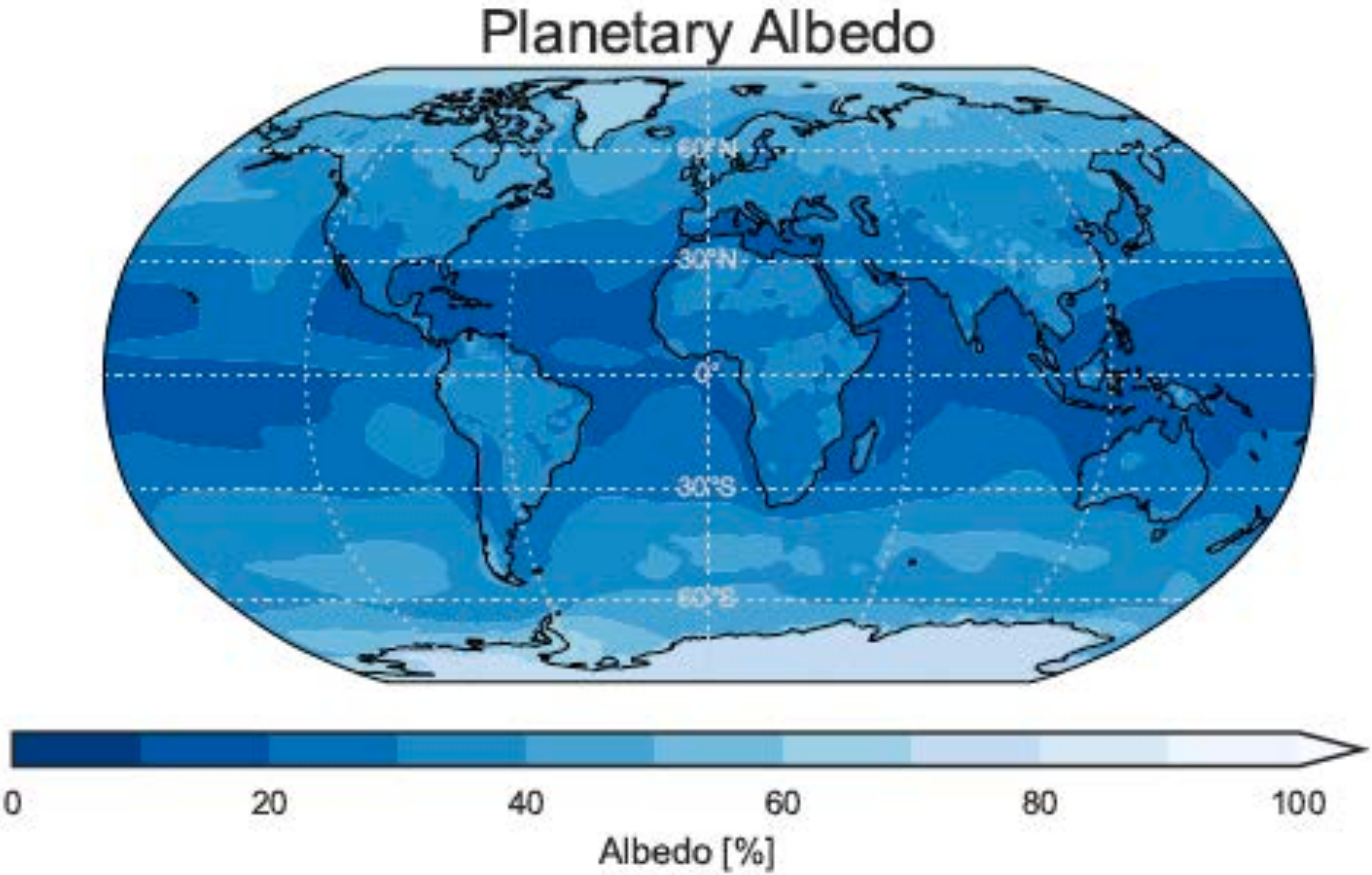
- Albedo ($\alpha_p$) : The ratio of reflected to incident solar energy
- On average,
  - $\alpha_p \approx 0.3$
- It is called as the planetary albedo.
- Solar radiation absorbed by the Earth:

$$(1-\alpha_p)\, S_0 \pi a^2 = 1.22 \times 10^{17} W$$

TABLE 2.2. Albedos for different surfaces. Note that the albedo of clouds is highly variable and depends on the type and form. See also the horizontal map of albedo shown in Fig. 2.5.

| Type of surface | Albedo (%) |
|---|---|
| Ocean | 2–10 |
| Forest | 6–18 |
| Cities | 14–18 |
| Grass | 7–25 |
| Soil | 10–20 |
| Grassland | 16–20 |
| Desert (sand) | 35–45 |
| Ice | 20–70 |
| Cloud (thin, thick stratus) | 30, 60–70 |
| Snow (old) | 40–60 |
| Snow (fresh) | 75–95 |

Table 2.2, Marshall and Plumb (2008)


Planetary Albedo
60°N
30°N
0°
30°S
60°S
0
20
40
60
80
100
Albedo [%]

# 1. Planetary emission temperature

- The first law of thermodynamics : Energy is conserved.
- $$\frac{dT}{dt} = E_{in} - E_{out}$$
- We know $E_{in}$
- What is $E_{out}$?
- Following Stefan-Boltzmann law, the radiative energy that the Earth emits per unit area is

$$\sigma T_e^4$$

$$\sigma = 5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$$

# 1. Planetary emission temperature

- Total energy that the Earth emits:

$$E_{out} = 4\pi a^2 \sigma T_e^4$$

- Then by setting $E_{in} = E_{out}$, we can obtain the expression for the planetary emission temperature, $T_e$

$$T_e = \left[ \frac{S_0 \left(1 - \alpha_p\right)}{4\sigma} \right]^{1/4}$$

- If we use numbers we know, $T_e \approx 255 \text{ K}$

# 2. The atmospheric absorption

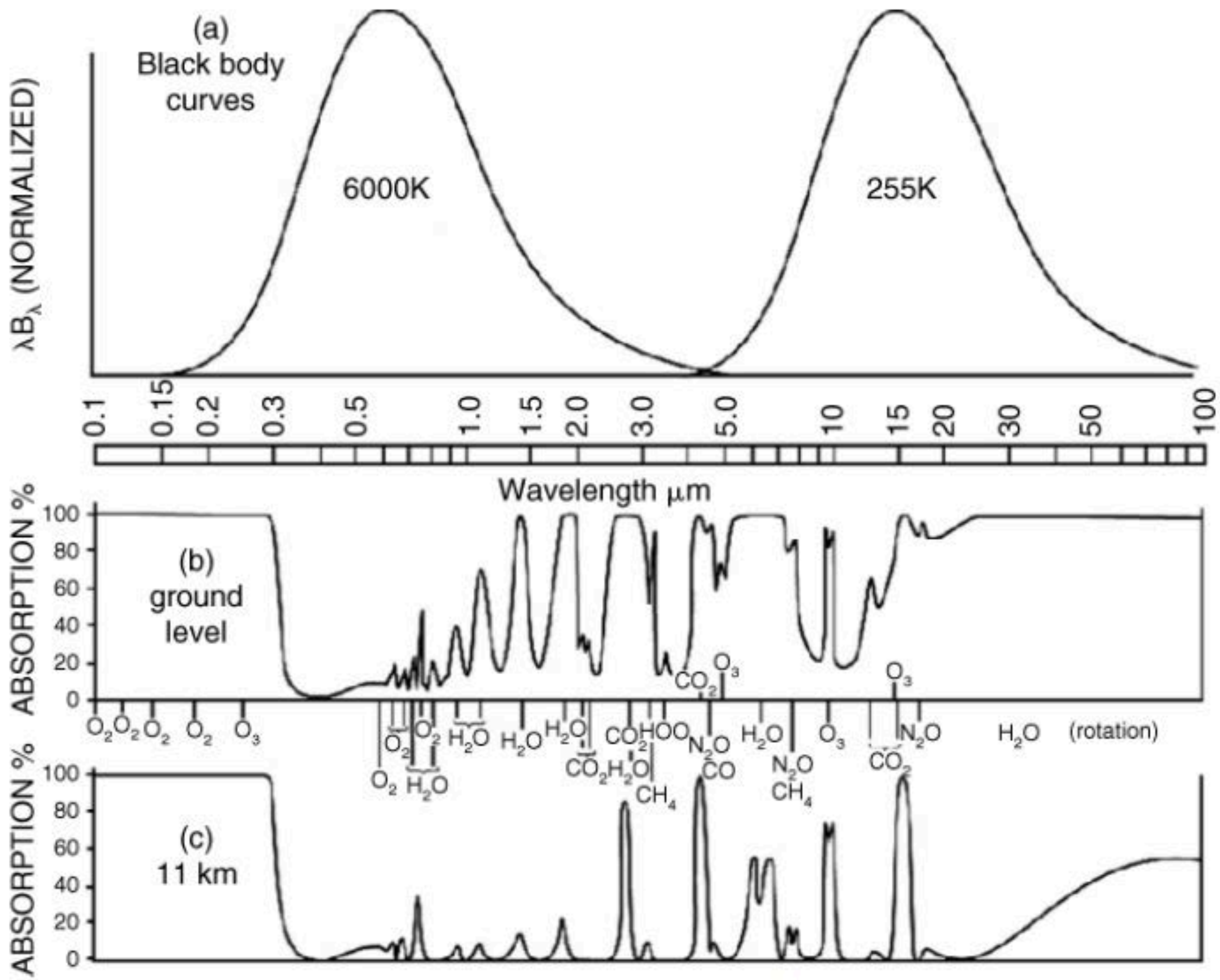



Figure 2.6, Marshall and Plumb (2008)

# 2. The atmospheric absorption

- The atmosphere is almost completely transparent in the visible spectrum.
- It is very opaque in the UV spectrum.
- The absorption of the IR spectrum by the atmosphere varies.
- Almost no contribution from $N_2$.
- $O_2$ absorbs in the UV (little solar energy) and near the IR spectrum.
- The absorption of the radiation occurs by triatomic molecules: $O_3$, $H_2O$ and $CO_2$

# 3. The greenhouse effect

- The emission temperature is too cold! $T_e \approx 255 \text{ K}$
- The atmosphere is not transparent to the IR.
- Much of the radiation from the surface will be absorbed by, mainly $H_2O$ and comes back to the surface.
- Hence, the surface gets both solar radiation and longwave radiation from the atmosphere and is warmer than $T_e$.
- This is known as *the greenhouse effect*.

# Contribution

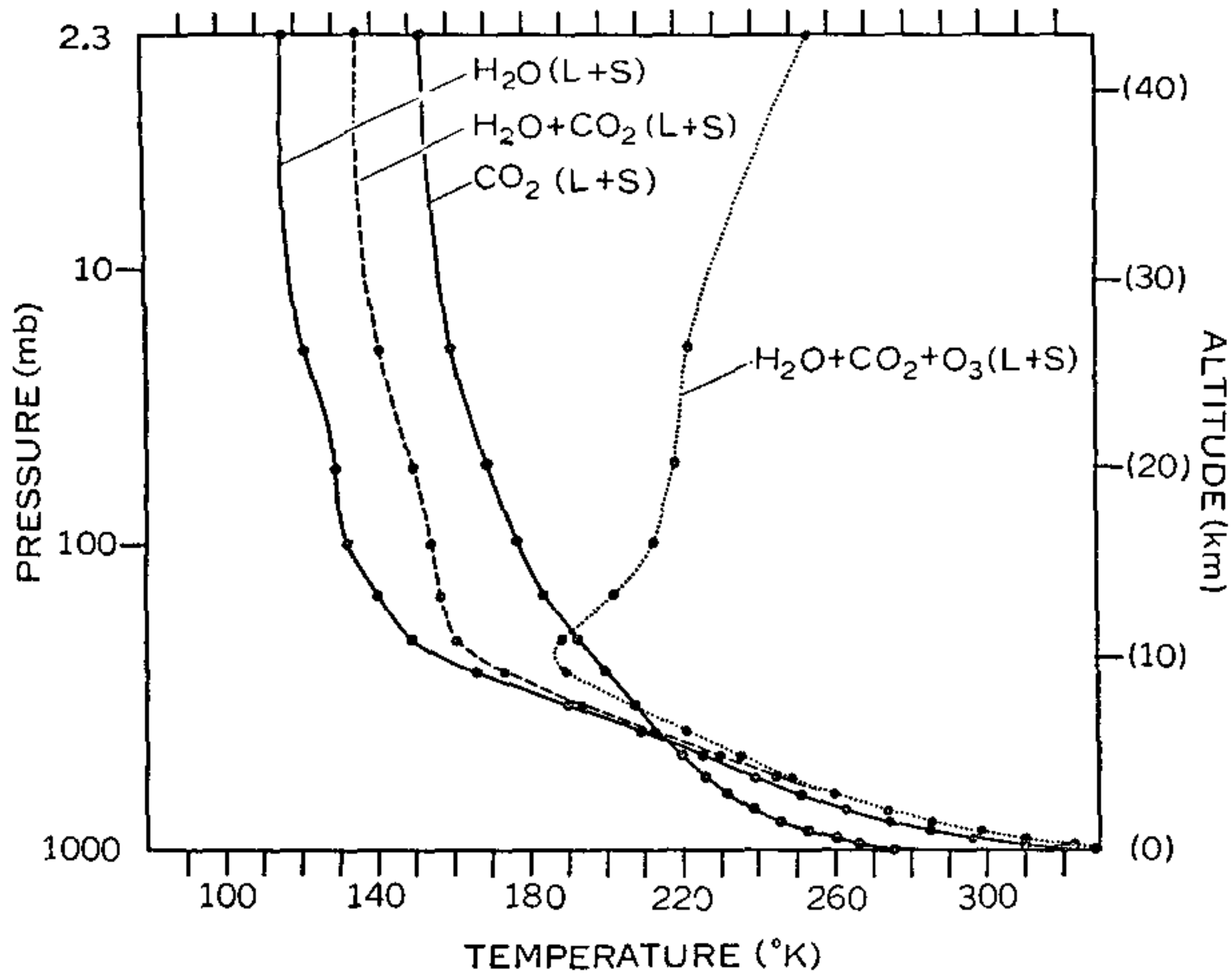



Figure 6a, Manabe and Stricker (1964)

# The Nobel prize in physics, 2021

Ill. Niklas Elmehed © Nobel Prize Outreach

Syukuro Manabe

Prize share: 1/4



Ill. Niklas Elmehed © Nobel Prize Outreach

Klaus Hasselmann

Prize share: 1/4



Ill. Niklas Elmehed © Nobel Prize Outreach

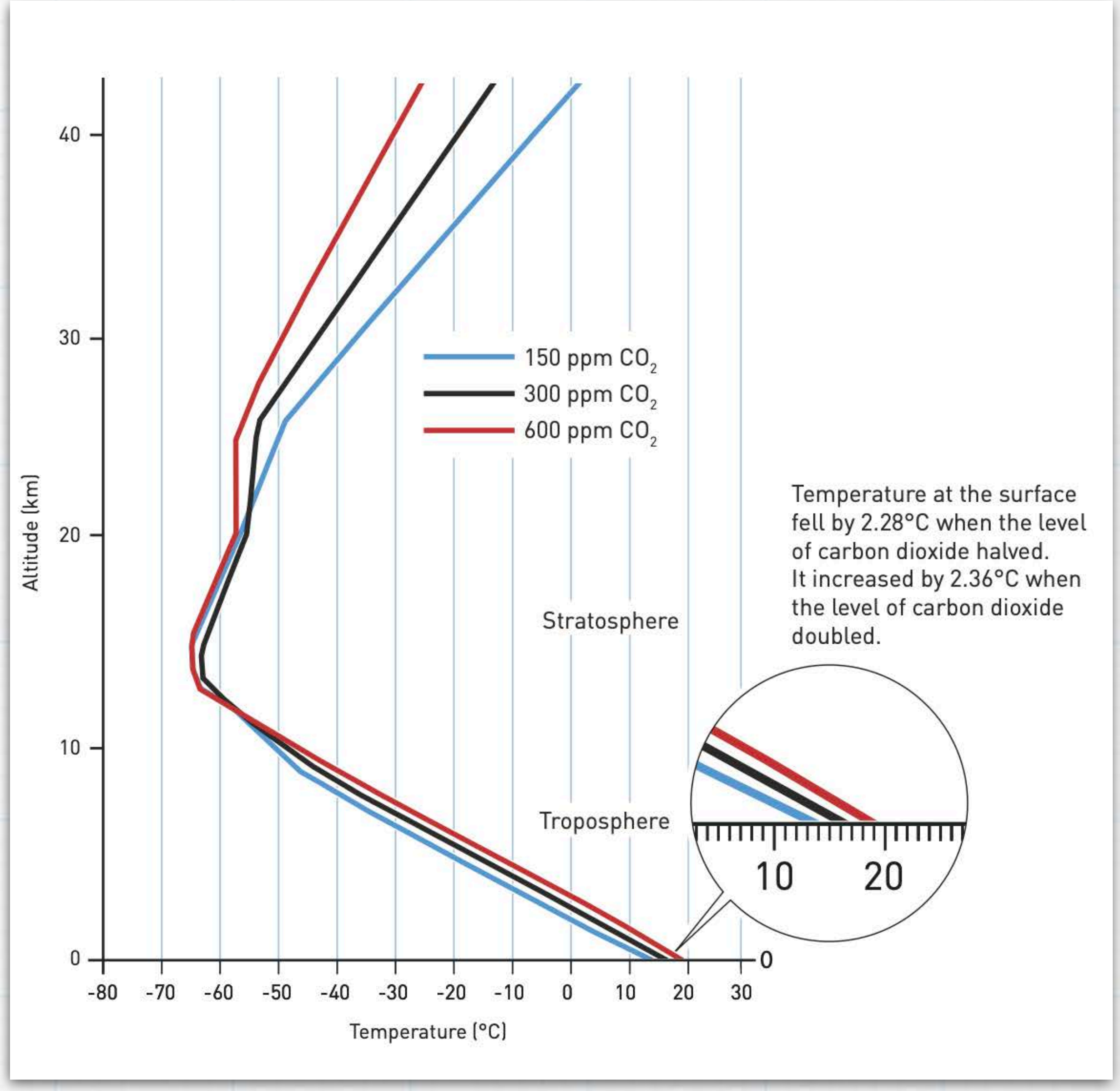
Giorgio Parisi

Prize share: 1/2

# Prediction of $\Delta T$ in response to $\Delta CO_2$

- **Manabe and Wetherald (1967)**
  - Nailed that $CO_2$ is important GHG.
  - 150 ppm v.s. 300 ppm v.s. 600 ppm

    – 2.28°C 2.36°C
  - The increased level of $CO_2$ increases the surface temperature while decreases that in the stratosphere.
  - This is Suki's favorite paper.
  - Suki didn't realize this paper would draw lots of interest.



# Hydrostatic Balance

- Now, the mass of the cylinder is

$$M = \rho \delta A \delta z$$

- If this cylinder is not accelerating, the net force should be zero!
  - Gravitational force ($F_g$)
  - Pressure force at the top ($F_T$)
  - Pressure force at the bottom ($F_B$)

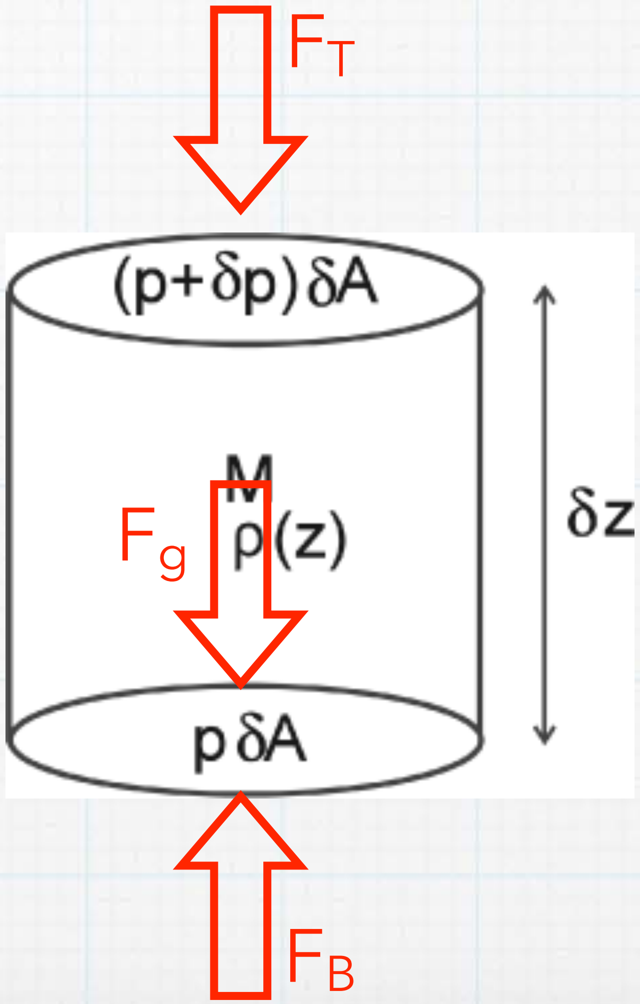



Figure 3.5, Marshall and Plumb (2008)

# Hydrostatic Balance

- $F_g = -gM = -g\rho\delta A\delta z$
- $F_T = -(p + \delta p)\delta A$
- $F_B = p\delta A$
- $F_g + F_T + F_B = \delta p + g\rho\delta z = 0$
- The equation of hydrostatic balance:

$$\frac{\partial p}{\partial z} + g\rho = 0$$

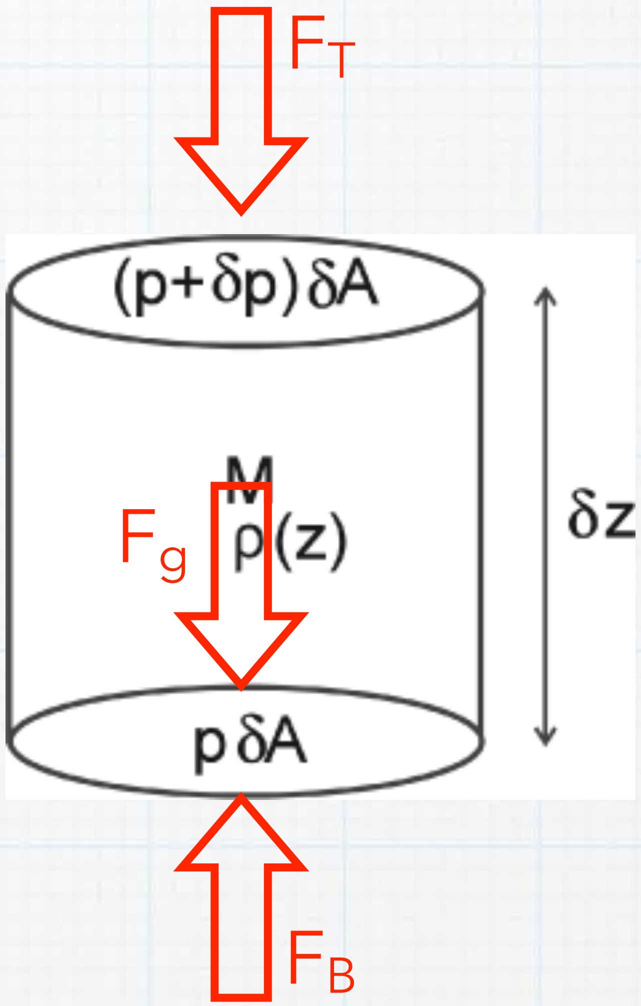



Figure 3.5, Marshall and Plumb (2008)

# Potential temperature

- For **dry air,** the rate of temperature decrease is constant:

$$\frac{dT}{dz} = -\frac{g}{c_p} = -\Gamma_d$$

- Using hydrostatic balance, we can rewrite this as

$$c_p dT = -g dz = \frac{1}{\rho} dp$$

- Then, using the perfect gas law, this equation becomes

$$\frac{dT}{T} = \frac{R}{c_p}\frac{dp}{p} = \kappa \frac{dp}{p}$$

# Potential temperature

- Further, this PDE can be arranged as

$$d \ln T = \kappa d \ln p$$

- And we get this relationship: $\dfrac{T}{p^\kappa} = const.$

- It means that T has to go down as p decreases, or vice versa.

- If we integrate the first equation from $p=p_0$ to $p=p$,

$$T(p_0) = T(p) \left( \frac{p_0}{p} \right)^\kappa$$

# Potential temperature

- Let's replace $T(p_0)$ with $\theta$.

$$\theta = T(p)\left(\frac{p_0}{p}\right)^{\kappa}$$

- $\theta$ is called as potential temperature, and it represents the temperature at $p=p_0$. (conventionally, $p_0$ is 1000 mb.)
- We introduced potential temperature to get a quantity that does not rely on height (or p), but there is p in that equation. So we failed?

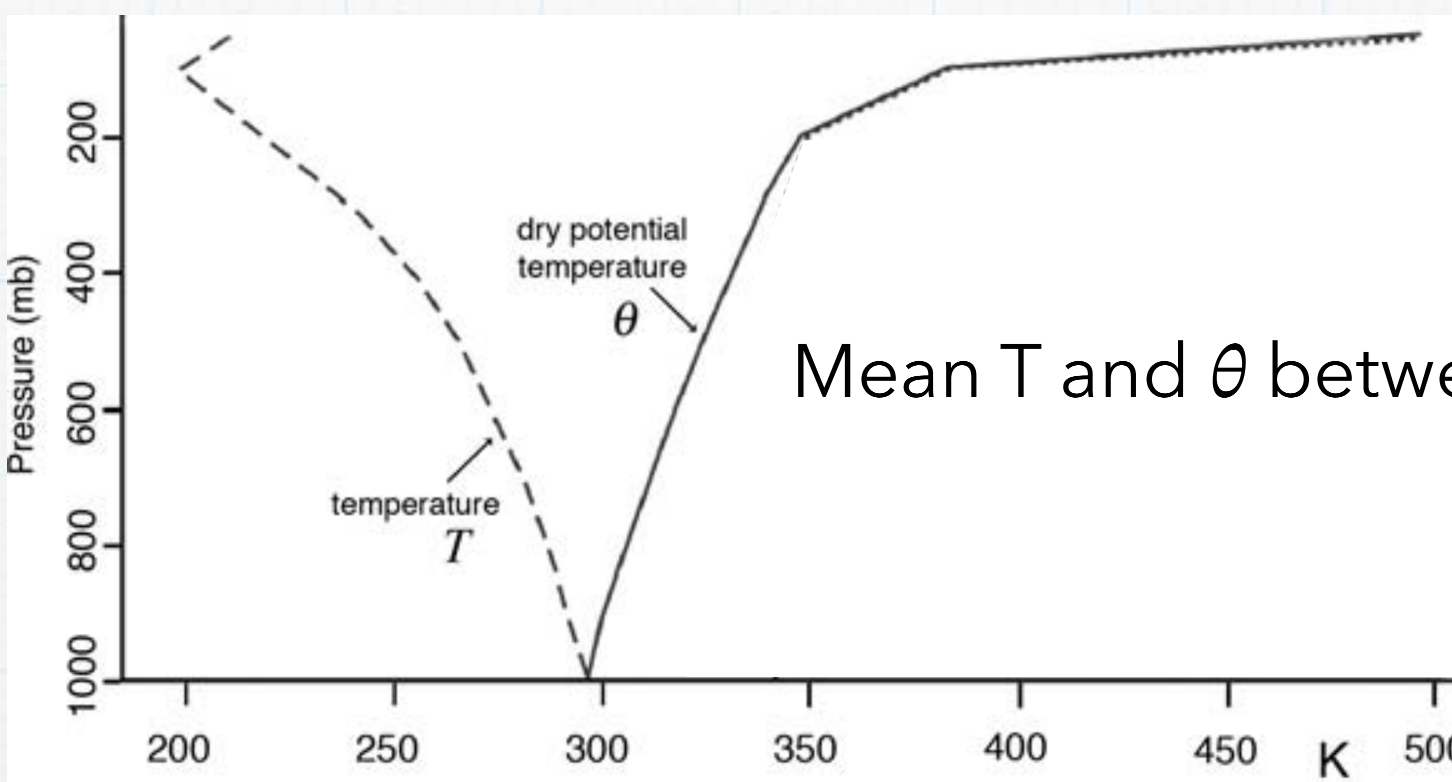
# Potential temperature

- If $\theta$ does not depend on p, then d$\theta$/dp should be zero.

$$\frac{d\theta}{dp} = \frac{dT}{dp}\left(\frac{p_0}{p}\right)^{\kappa} - \kappa\frac{T}{p}\left(\frac{p_0}{p}\right)^{\kappa} = 0$$

- T and $\theta$ have to converge at p=1000 mb.

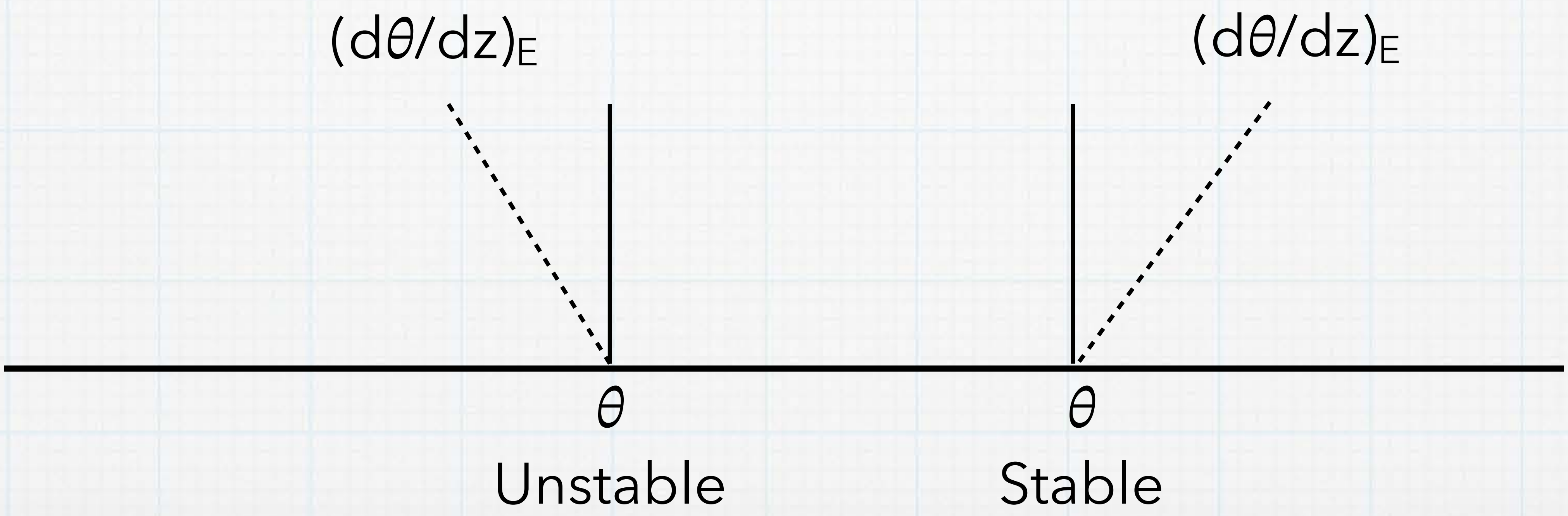
Figure 4.9, Marshall and Plumb (2008)




Mean T and $\theta$ between 30S and 30N

# Potential temperature and stability

- Stability using potential temperature, $\theta$
  - Unstable if $(d\theta/dz)_E < 0$
  - Neutral if $(d\theta/dz)_E = 0$
  - Stable if $(d\theta/dz)_E > 0$



# Moist convection : humidity

- We need a measure for how wet the air is.
- **Specific humidity (q)** : the mass of water vapor to the mass of air per unit volume

The mass of water vapor

$$q = \frac{\rho_v}{\rho} = \frac{\rho_v}{\rho_d + \rho_v}$$

The total mass of air = the mass of water vapor + the mass of dry a

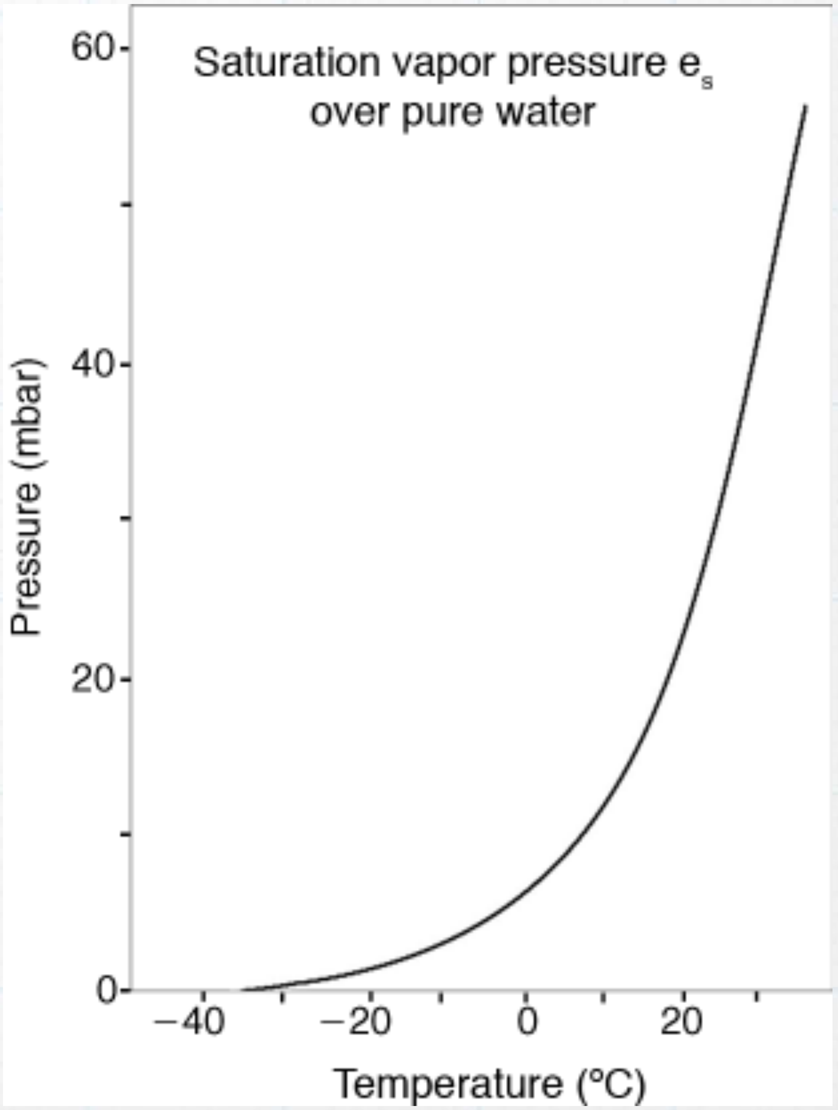
# Moist convection : humidity

- We need a measure for how wet the air is.
- **Saturation-specific humidity ($q_*$)** : the specific humidity at which saturation occurs

The mass of water vapor at saturation

$$q_* = \frac{\rho_{v,*}}{\rho} = \frac{e_s/R_vT}{p/RT} = \left(\frac{R}{R_v}\right)\frac{e_s}{p}$$

$$q_* = q_*(p,T)$$




$$e_s = A\exp(\beta T)$$

# Moist convection : humidity

- **Relative humidity** : the ratio of the specific humidity to the saturation specific humidity

$$U = \frac{q}{q_*} \times 100\%$$

- The surface has higher humidity than aloft (relative humidity is close to 80%).
- Raise humid air..
  - Both p and T decrease, and $q_*$ goes up? Or down?
  - How about q?
  - What happens if $q = q_*$?

# Saturated adiabatic lapse rate

- Using saturation specific humidity and saturated partial pressure of water vapor, one can convert the equation in the previous slide to

$$-\frac{dT}{dz} = \Gamma_s = \Gamma_d \underbrace{\left[\frac{1 + Lq_*/RT}{1 + \beta L q_*/c_p}\right]}_{<1}$$

↓ Saturated adiabatic lapse rate

- $\Gamma_s < \Gamma_d$
- $\Gamma_s$ is a function of both p and T, and is $3\ \mathrm{K/km} < \Gamma_s < 10\ \mathrm{K/km}$

# Saturated adiabatic lapse rate

- What would be $\Gamma_s$ in tropical lower troposphere?
- What would be $\Gamma_s$ in the upper troposphere?
- Condensation releases latent heat. => air becomes more buoyant
- The presence of water vapor destabilizes the atmosphere.

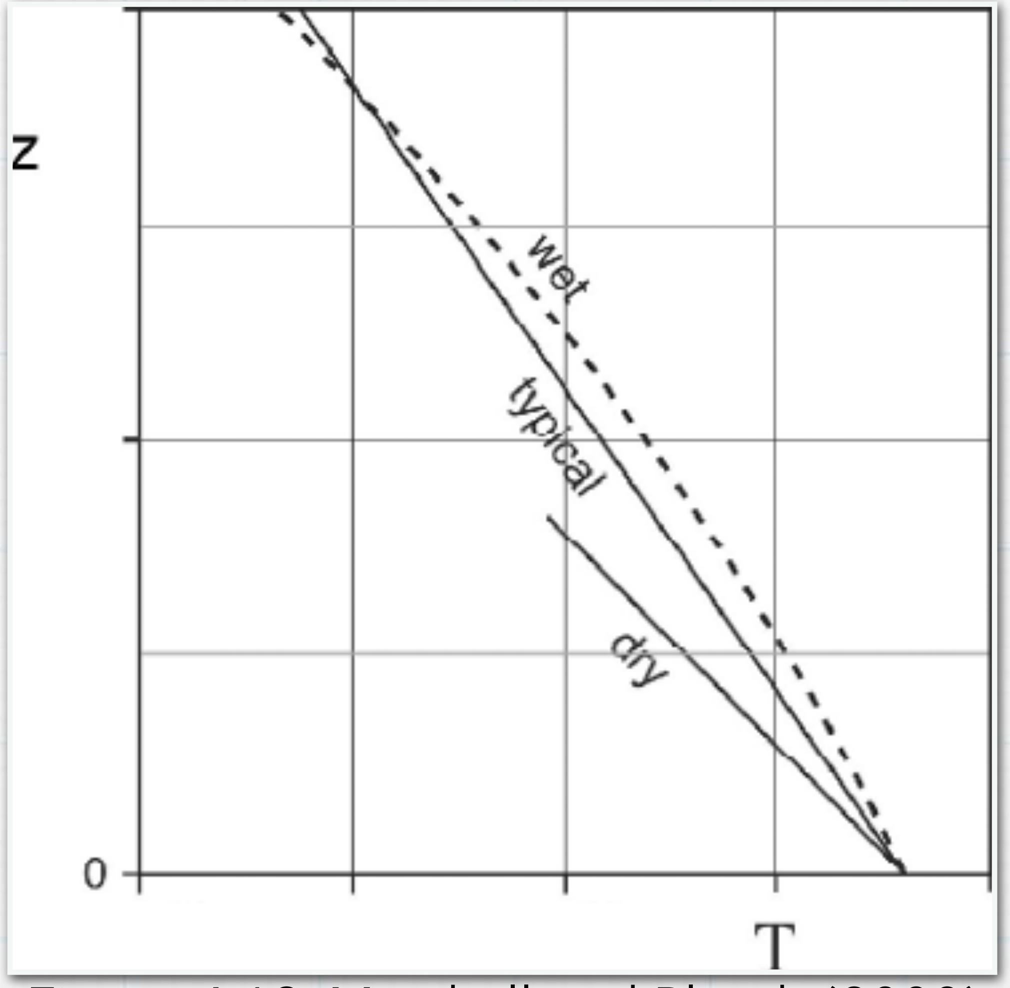


Figure 4.19, Marshall and Plumb (2008)

# Stability, AGAIN!