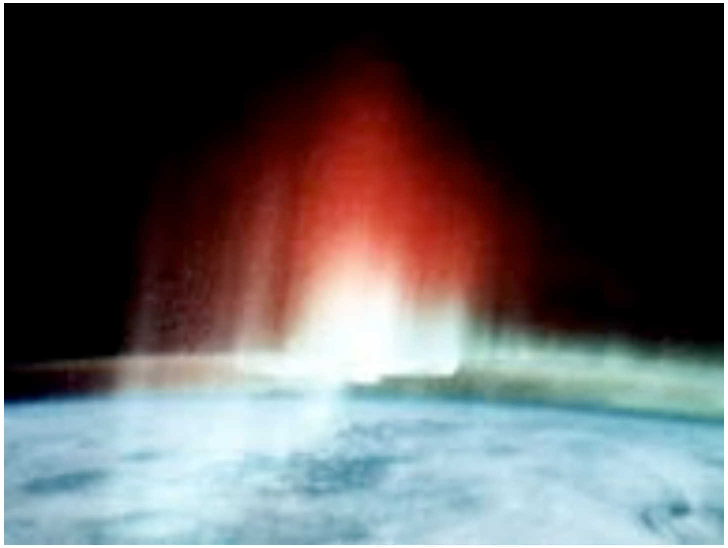
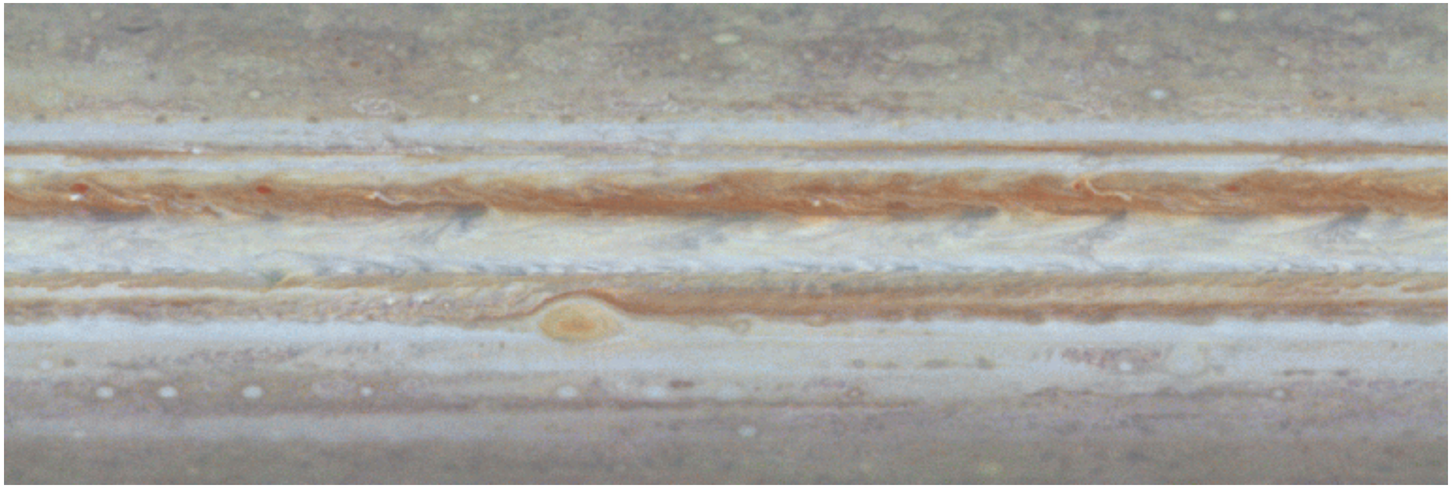
# Planetary Atmospheres

- Structure
- Composition
- Clouds
- Photochemistry
- Meteorology
- Atmospheric Escape





# Structure

## Generalized Hydrostatic Equilibrium

$$P(z) = P(0)e^{-\int_0^z dr/H(r)} \qquad \rho(z) = \rho(0)e^{-\int_0^z dr/H^*(r)}$$

## Generalized Pressure Scale Height

$$H(z) = \frac{kT(z)}{g_p(z)\mu_a(z)m_{amu}}$$

## Generalized Density Scale Height

$$\frac{1}{H^*(z)} = \frac{1}{T(z)}\frac{dT(z)}{dz} + \frac{g_p(z)\mu_a(z)m_{amu}}{kT(z)}$$

# Structure

Note: For an Isothermal Atmosphere

(or region of an atmosphere):

$$H(z) = H^*(z)$$

Since $$\frac{dT(z)}{dz} = 0$$

Remember that $$g_p(z) = \frac{GM_p}{r^2} = \frac{GM_p}{\left(R_p + z\right)^2}$$

* So at small altitudes $r \Rightarrow R_P$ *and* $g_p(z) \cong g_p(R_p)$

# Structure

Most planets have near-surface scale heights ranging between ~10-25 km due to the similar ratios of

$T/(g_p \mu_a)$

| | Venus | Earth | Mars | Jupiter | Saturn | Uranus | Neptune |
|---|---|---|---|---|---|---|---|
| $T_{surf}$ (K) | 737 | 288 | 215 | 165* | 135* | 76* | 72* |
| Bond Albedo | 0.75 | 0.31 | 0.25 | 0.34 | 0.34 | 0.29 | 0.31 |
| $H$ (km) | 16 | 8.5 | 11 | 24 | 47 | 25 | 23 |

* Temperature at 1 bar pressure

# Structure

Of course, temperature actually does vary with height

If a packet of gas rises rapidly (adiabatic), then it will expand and, as a result, cool

Work done in expanding = work done in cooling

$$VdP = \frac{m_{gm}}{\rho} dP \qquad\qquad C_p dT$$

$m_{gm}$ is the mass of one mole, $\rho$ is the density of the gas

$C_p$ is the specific heat capacity of the gas at constant pressure

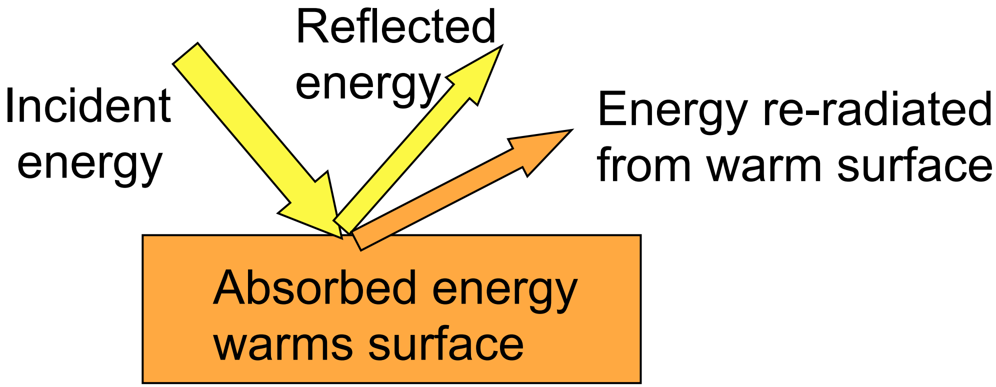
Combining these two equations with hydrostatic equilibrium, we get the dry adiabatic lapse rate:

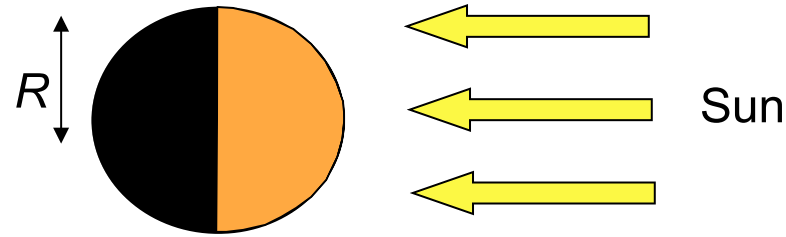
$$\frac{dT}{dz} = \frac{m_{gm} g_p}{C_p} = \frac{g_p}{c_p}$$

* On Earth, the lapse rate is about 10 K/km

# Thermal Structure: Surface

## What determines a planet's surface temperature?




$$P_{in} = (1 - A_b)\pi R^2 \frac{F_\odot}{r_{\odot AU}^2}$$

$$P_{out} = 4\pi R^2 \varepsilon \sigma T^4$$

$A_b$ is Bond albedo, $F_\odot$ is solar flux at Earth's distance, $r_\odot$ is distance of planet to Sun, $\varepsilon$ is emissivity, $\sigma$ is Stefan's constant ($5.67\mathrm{x}10^{-8}\ \mathrm{Wm^{-2}K^{-4}}$)

## Balancing energy in and energy out yields:

$$T_{eq} = \left( \frac{F_\odot}{r_{\odot AU}^2} \frac{(1 - A_b)}{4\varepsilon\sigma} \right)^{1/4}$$

# Thermal Structure: Surface

- Solar constant $F_{\odot}$ =1300 Wm$^{-2}$
- Earth (Bond) albedo $A_b$=0.3, $\varepsilon$=0.9
- Equilibrium temperature = 263 K
- How reasonable is this value?

| Body | Mercury | Venus | Earth | Mars |
|---|---|---|---|---|
| $A_b$ | 0.12 | 0.75 | 0.3 | 0.25 |
| $T_{eq}$ | 446 | 238 | 263 | 216 |
| Actual $T$ | 100-725 | 737 | 288 | 215 |

- How to explain the discrepancies?
- Has the Sun's energy stayed constant with time?

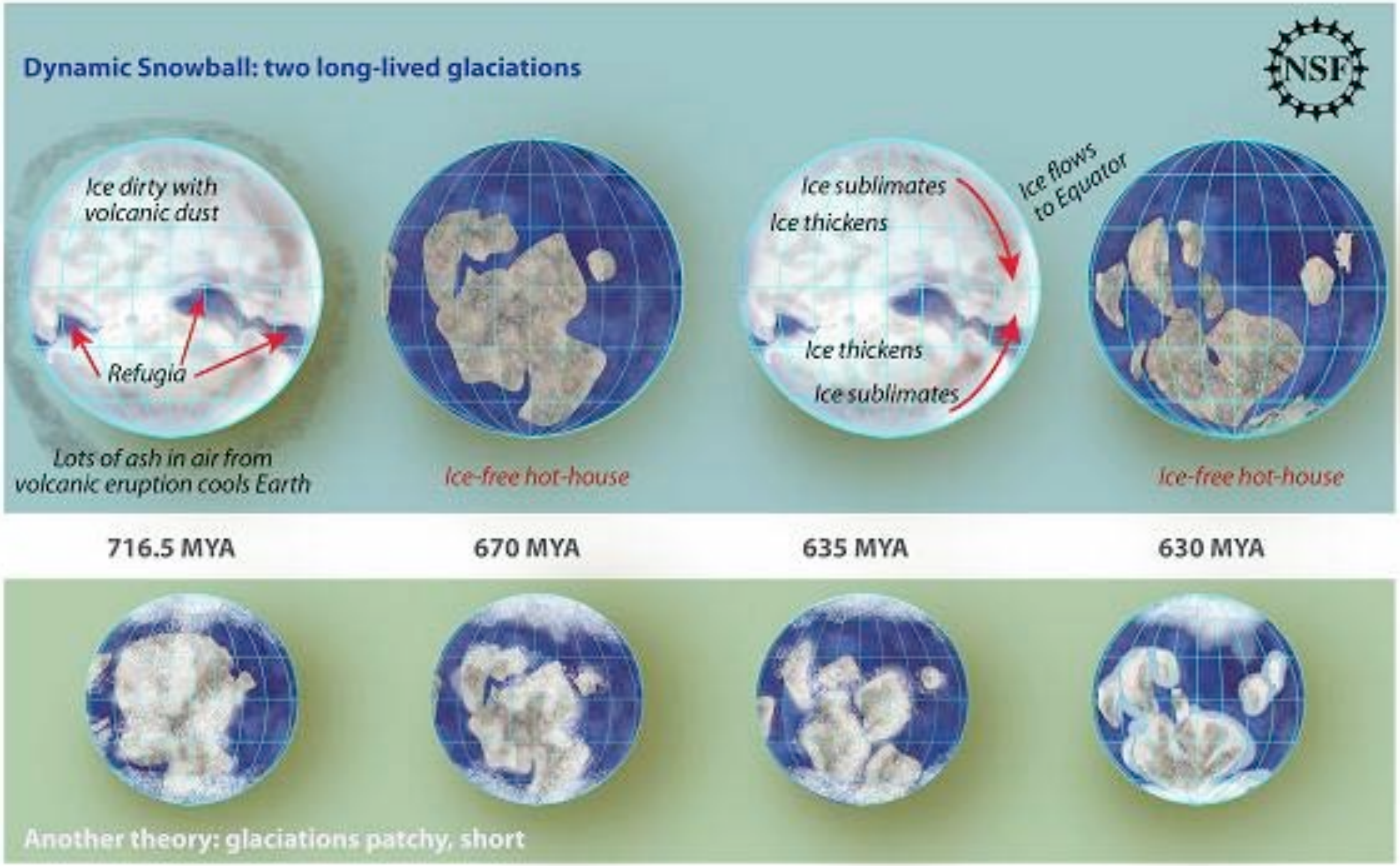
# Thermal Structure: Greenhouse Effect

- Atmosphere is more or less transparent to radiation (photons) depending on wavelength – opacity
- Opacity is low at visible wavelengths, high at infrared wavelengths due to absorbers like water vapor, $CO_2$
- Incoming light (visible) passes through atmosphere with little absorption
- Outgoing light is infrared (since the surface temperature is lower) and is absorbed by atmosphere
- So atmosphere heats up
- Venus suffered from a *runaway* greenhouse effect – surface temperature got so high that water was unstable, so no $CO_2$ could dissolve and form carbonates as on Earth

# Thermal Structure: Albedo Effects

- Fraction of energy reflected (not absorbed) by surface is given by the albedo $A$ ($0<A<1$)
- Coal dust has a low albedo, ice a high one
- The albedo can have an important effect on surface temperature
- E.g. ice caps grow, albedo increases, more heat is reflected, surface temperature drops, ice caps grow further . . . runaway effect!
- This mechanism is thought to have led to the Proterozoic Snowball Earth
- How might clouds affect planetary albedo?

# Recurring Snowball Earth?



# Atmospheric Thermal Structure

The atmospheric temperature profile is governed by the efficiency of energy transport, which largely depends on optical depth, $\tau_\nu$ . Remember that heating by solar radiation is a 'top-down' process.

Optical depth (or transparency) is determined by physical and chemical processes in the atmosphere and can change in time and in altitude.

# Atmospheric Thermal Structure

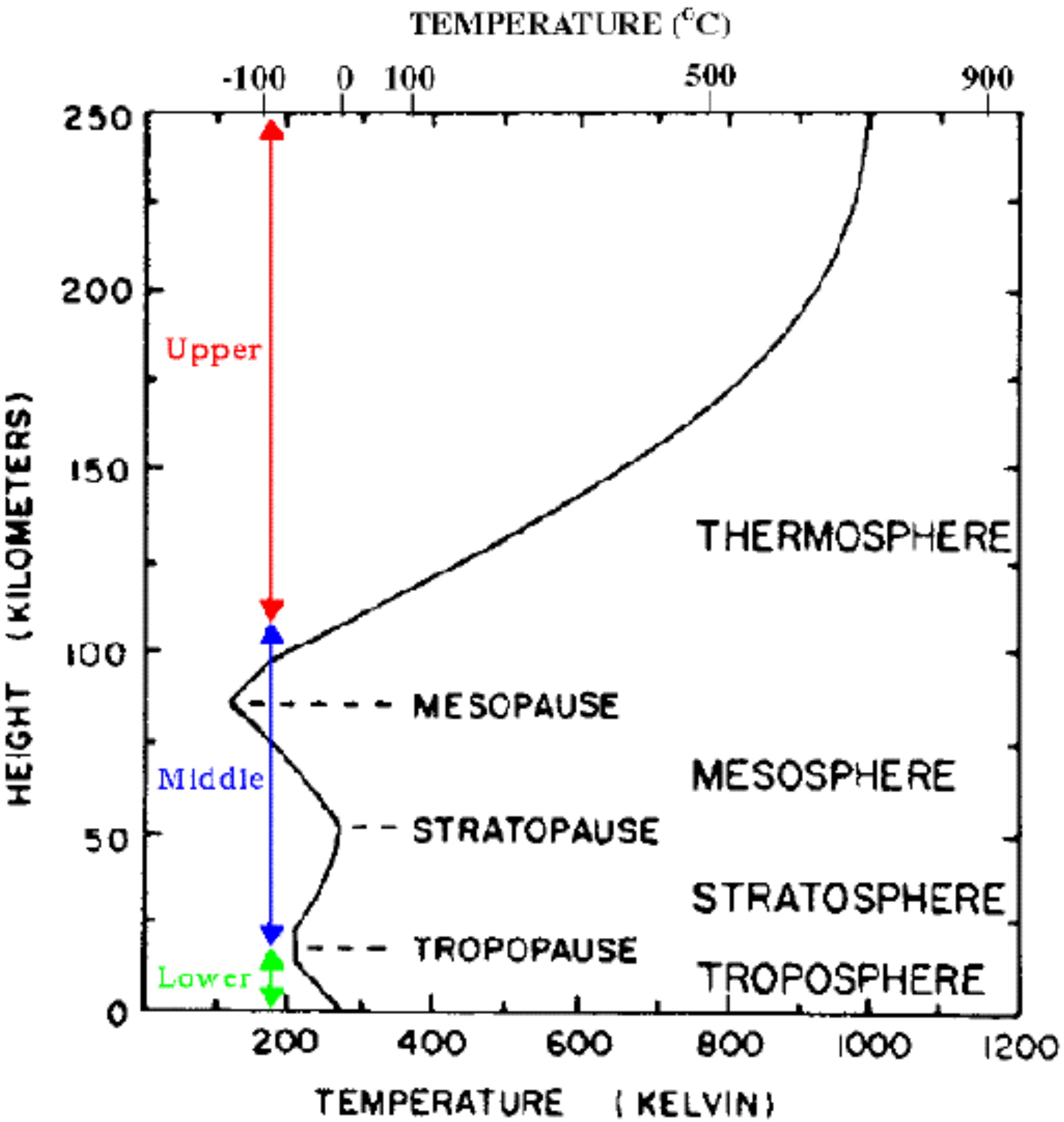
Other factors to consider:

Clouds can change the albedo, the optical depth, and the local temperature (via release/ absorption of latent heat).

Surface variations/composition can affect albedo and surface temperatures depend on the thermal properties of materials and their chemical interactions with the atmosphere

Geologic processes such as volcanism can greatly impact the composition, as well as chemistry and albedo (via dust grains and aerosols) of the atmosphere.

# Atmospheric Thermal Structure




Troposphere: Where condensable gases form clouds. dT/dz < 0

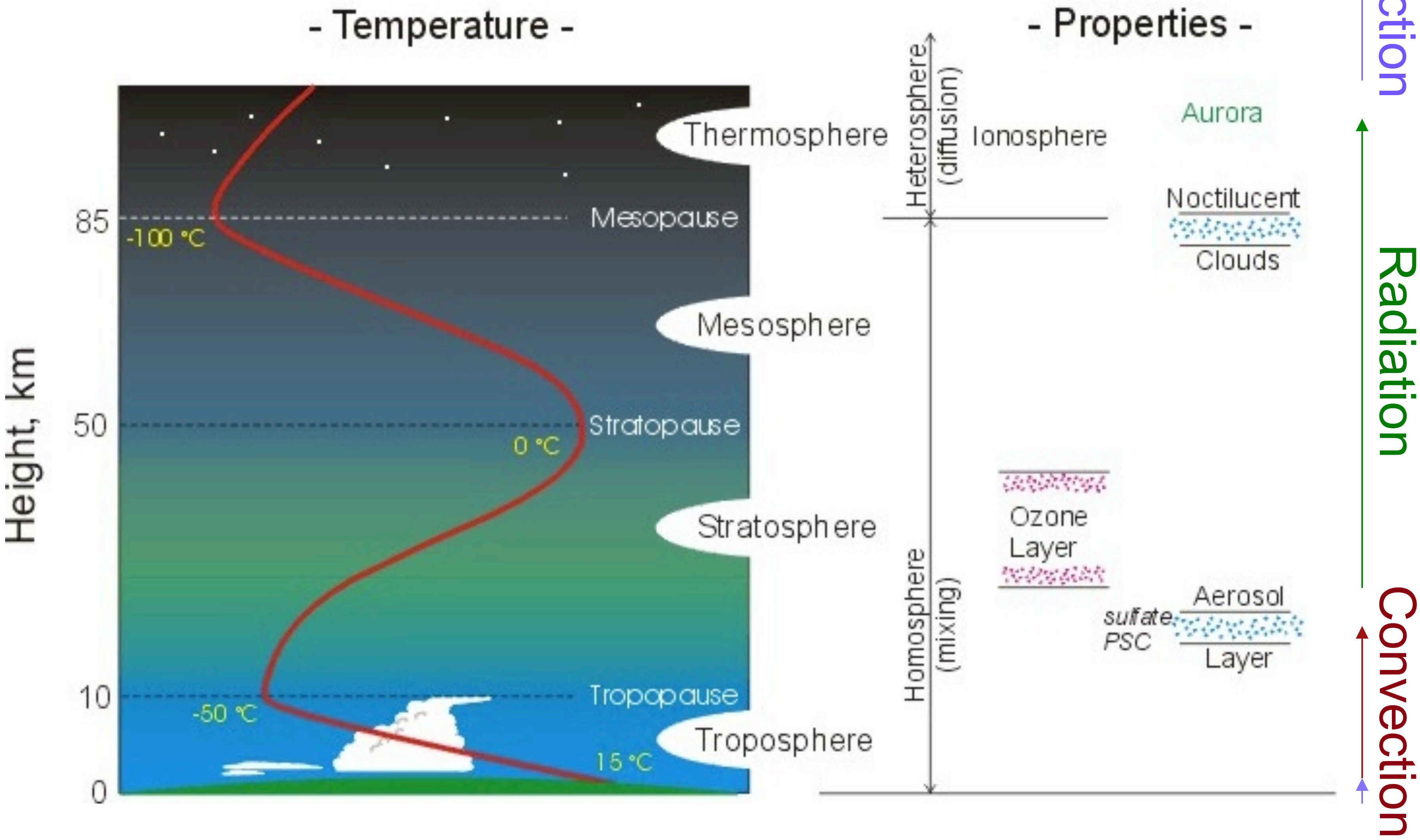
Stratosphere: dT/dz > 0

Mesosphere: dT/dz < 0

Thermosphere: dT/dz > 0

Exosphere: Roughly Isothermal

# Atmospheric Thermal Structure



# Atmospheric Thermal Structure

- Lower atmosphere (opaque) is dominantly heated from below and will be conductive or convective (adiabatic)
- Upper atmosphere intercepts solar radiation and re-radiates it
- There will be a temperature minimum where radiative cooling is most efficient (the tropopause)