12.003
Atmosphere, Ocean and Climate Dynamics

Lecture VI
Vertical Structure of the Atmosphere

Lecture VI Outline

1. Physical properties of dry air
2. Vertical distribution of temperature
3. Hydrostatic balance
4. Vertical distribution of pressure and density

Physical properties of dry air (p,ρ,T)

- Ideal gas law

$$p\,V = n\,R_g\,T = \frac{M}{m}\,R_g\,T$$

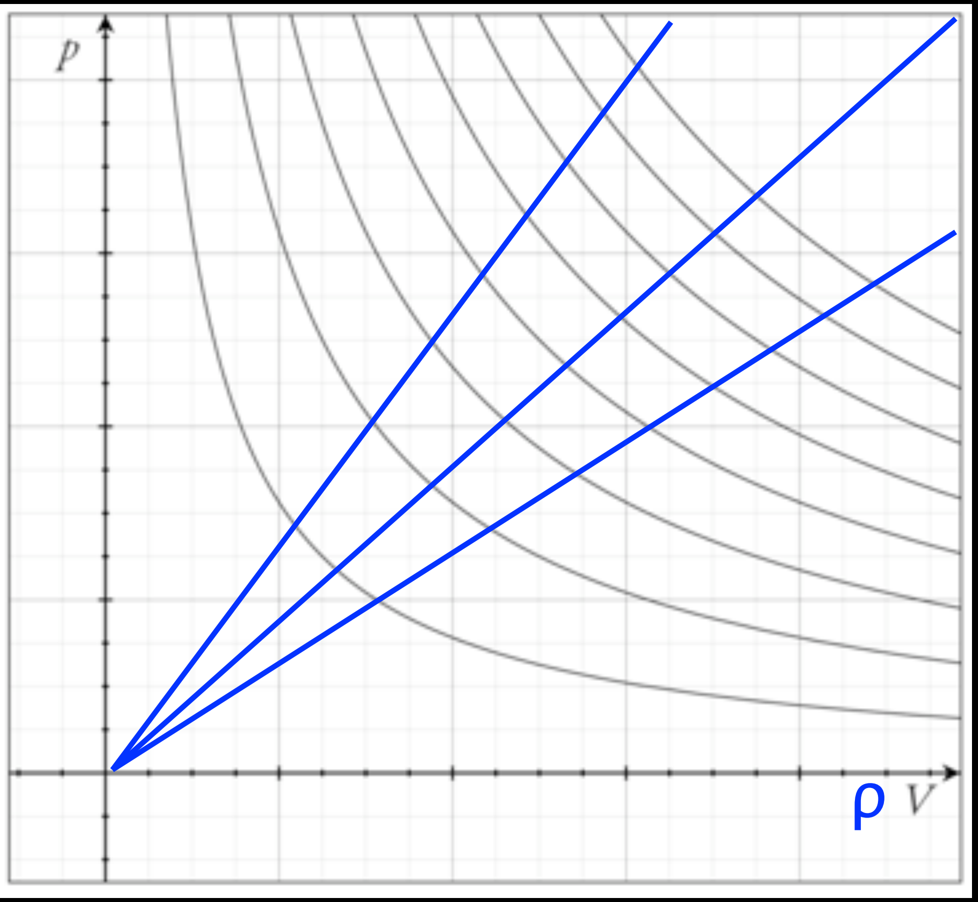
$$p = \rho\,\frac{R_g}{m}\,T = \rho\,R\,T$$

Universal gas constant

$$R_g = 8.314\ \mathrm{JK^{-1}mol^{-1}}$$

Gas constant of a specific gas

$$R = \frac{R_g}{m}$$

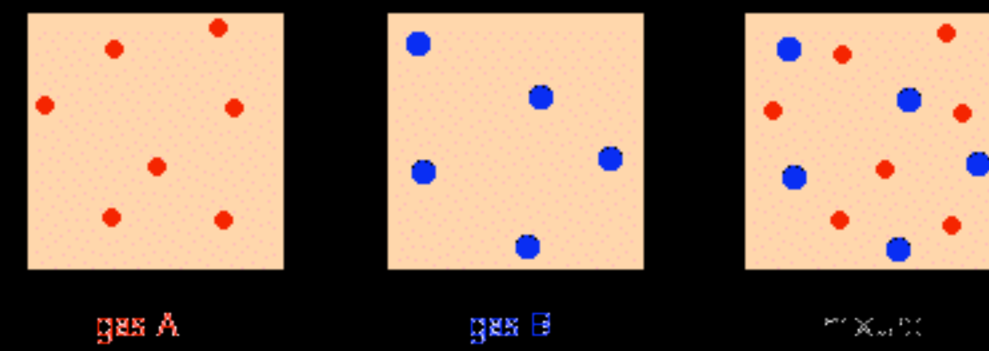


Animation

Physical properties of dry air

- Dalton's law of partial pressures for dry air

$$p\,V = \sum_i n_i\, R_g\, T = \sum_i \frac{M_i}{m_i}\, R_g\, T$$



$$p = R_g \frac{78\%\rho}{m_{N_2}}\, T + R_g \frac{21\%\rho}{m_{O_2}}\, T + \cdots = \rho \frac{R_g}{m_a}\, T = \rho\, R\, T$$

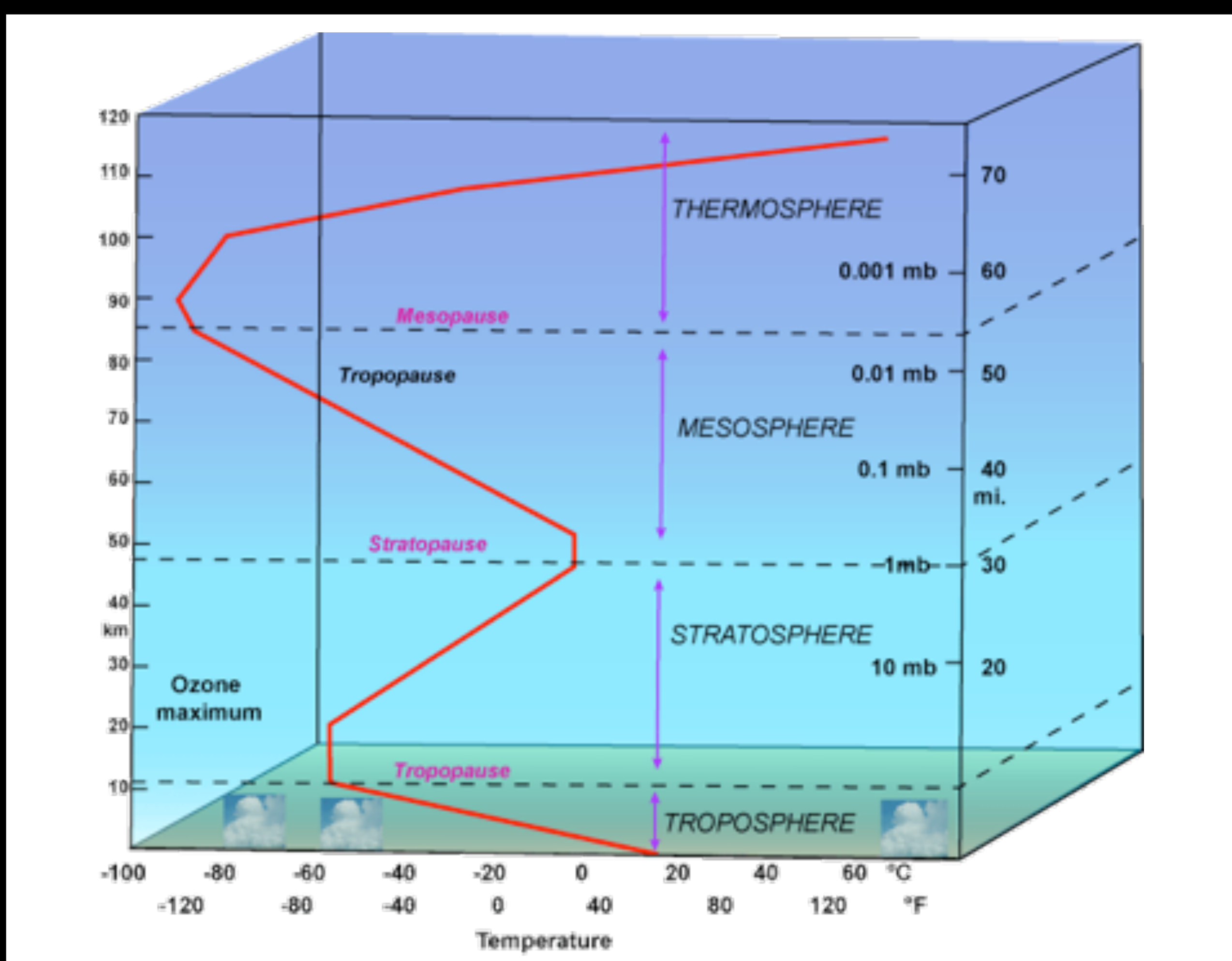
Components in Dry Air	Volume Ratio compared to Dry Air	Molecular Mass - *M* (kg/kmol)	Molecular Mass in Air
Oxygen	0.2095	32.00	6.704
Nitrogen	0.7809	28.02	21.88
Carbon Dioxide	0.0003	44.01	0.013
Hydrogen	0.0000005	2.02	0
Argon	0.00933	39.94	0.373
Neon	0.000018	20.18	0
Helium	0.000005	4.00	0
Krypton	0.000001	83.8	0
Xenon	$0.09\ 10^{-6}$	131.29	0
Total Molecular Mass of Air			28.97

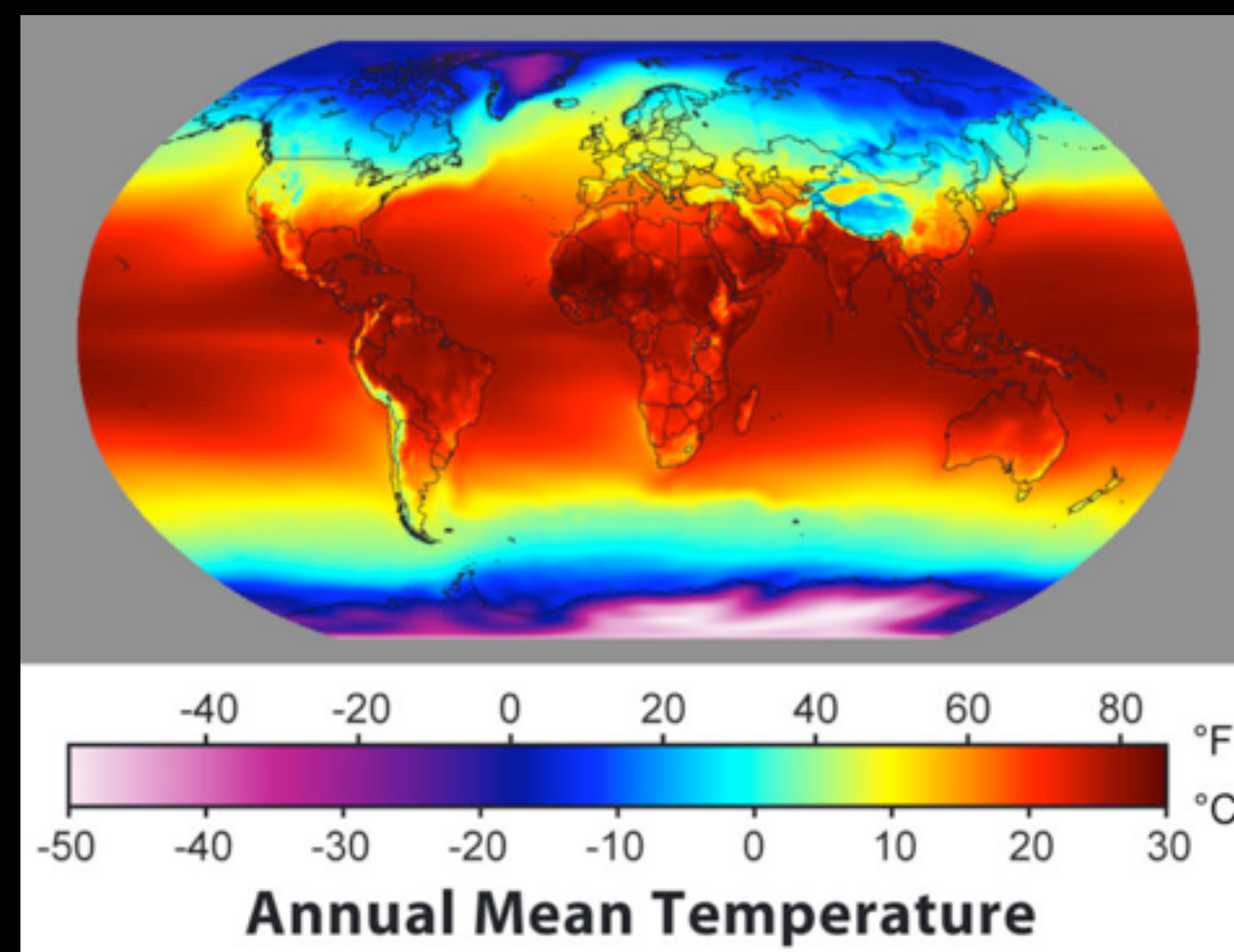
Dry air gas constant

$$R = \frac{R_g}{m_a} = 287\ \mathrm{Jkg^{-1}K^{-1}}$$

Atmospheric temperature variations

- Atmospheric temperature varies in horizontal and vertical
- Vertical variations are similar everywhere
  - ➡ pattern set by radiation budget (thermodynamics)
- Horizontal variations depend on height
  - ➡ pattern set atmospheric transport (dynamics)





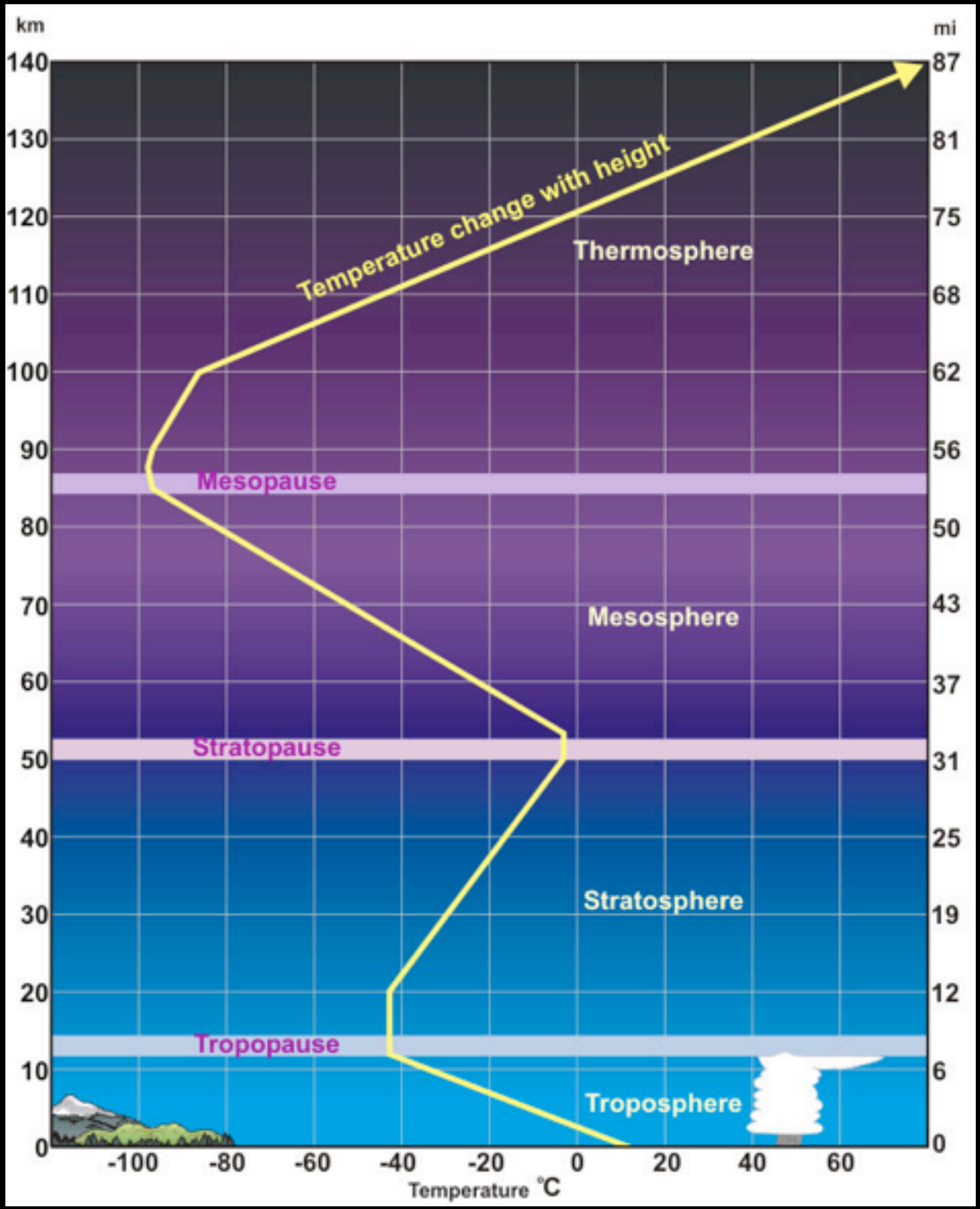
Vertical variations of temperature

Three hot spots:

1) thermopause =>

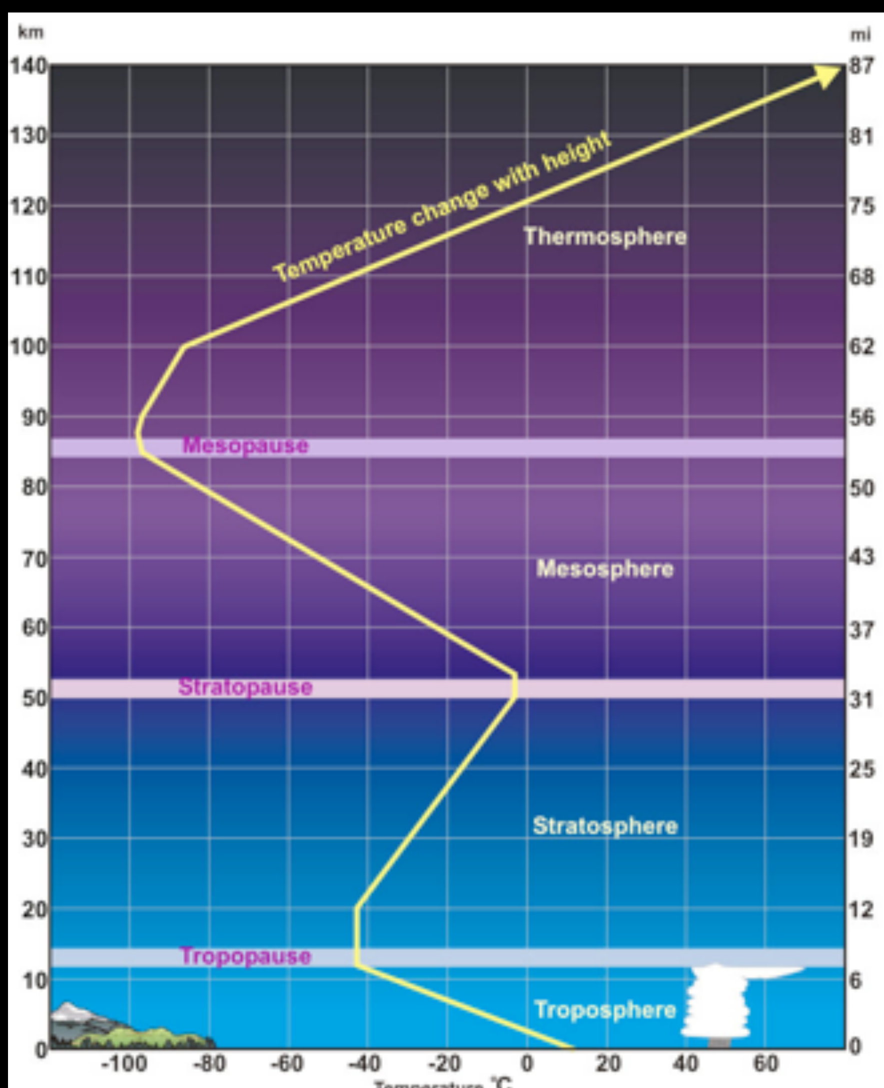
2) stratopause =>

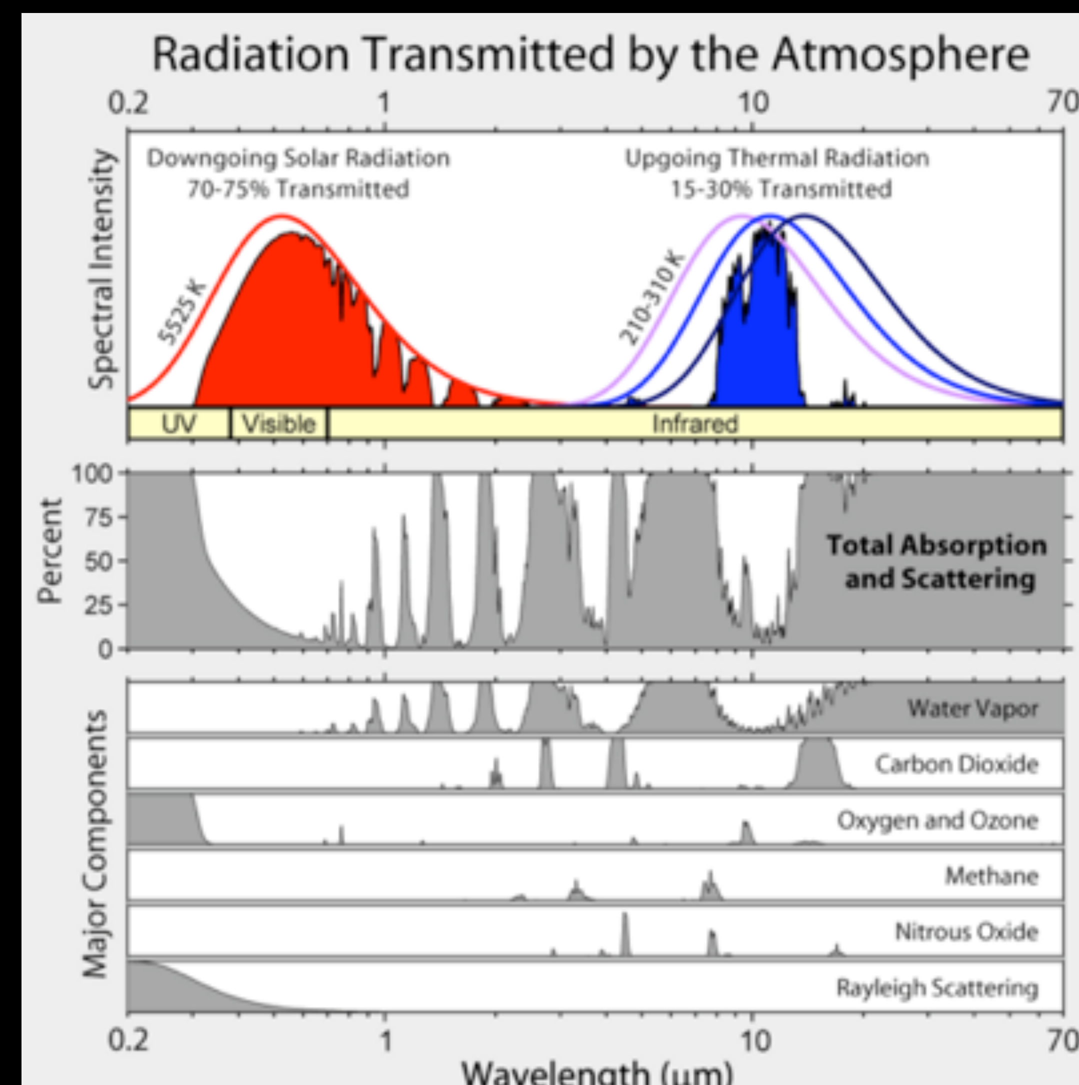
3) surface =>



Thermosphere

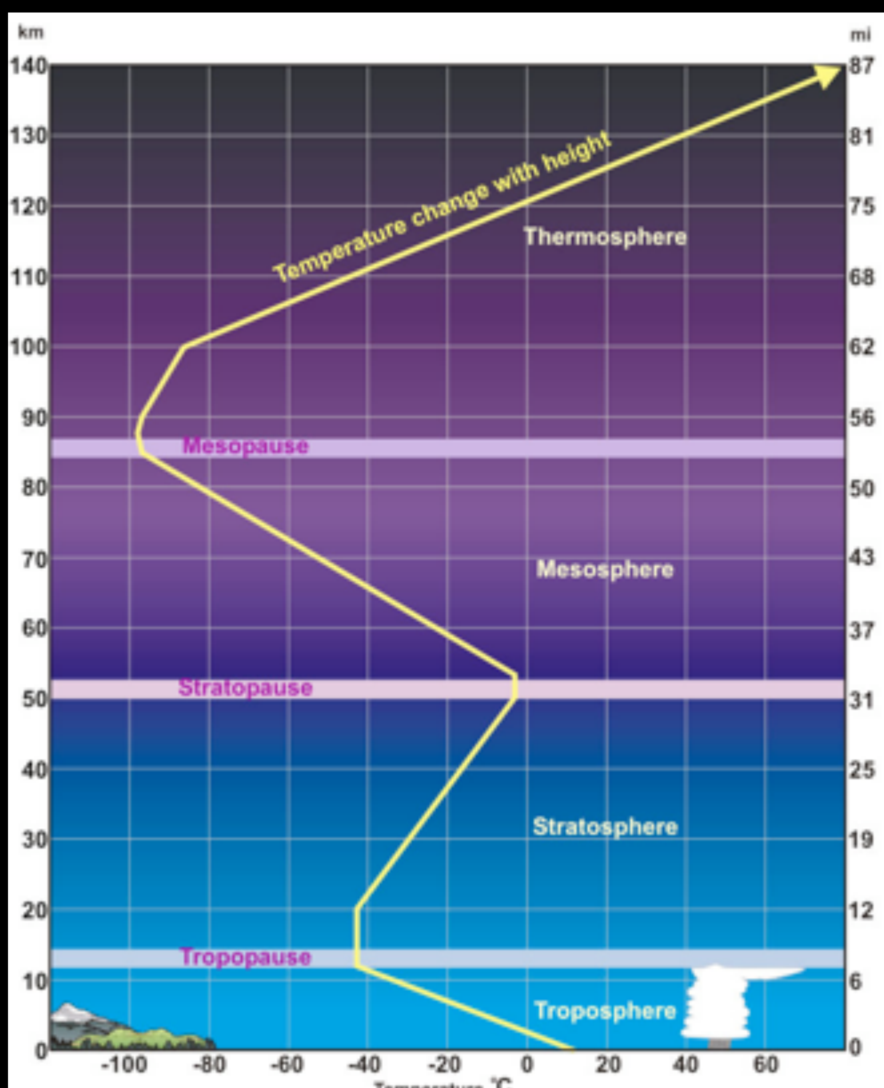
- Named from the Greek θερμός (heat) σφαιρα (sphere), begins about 90 km above Earth's surface
- Molecules are dissociated and electrons are ionized (no triatomic molecules)
- Absorbs high-energy UV (λ < 0.1 μm) - heated from above
- Temperature increases with height

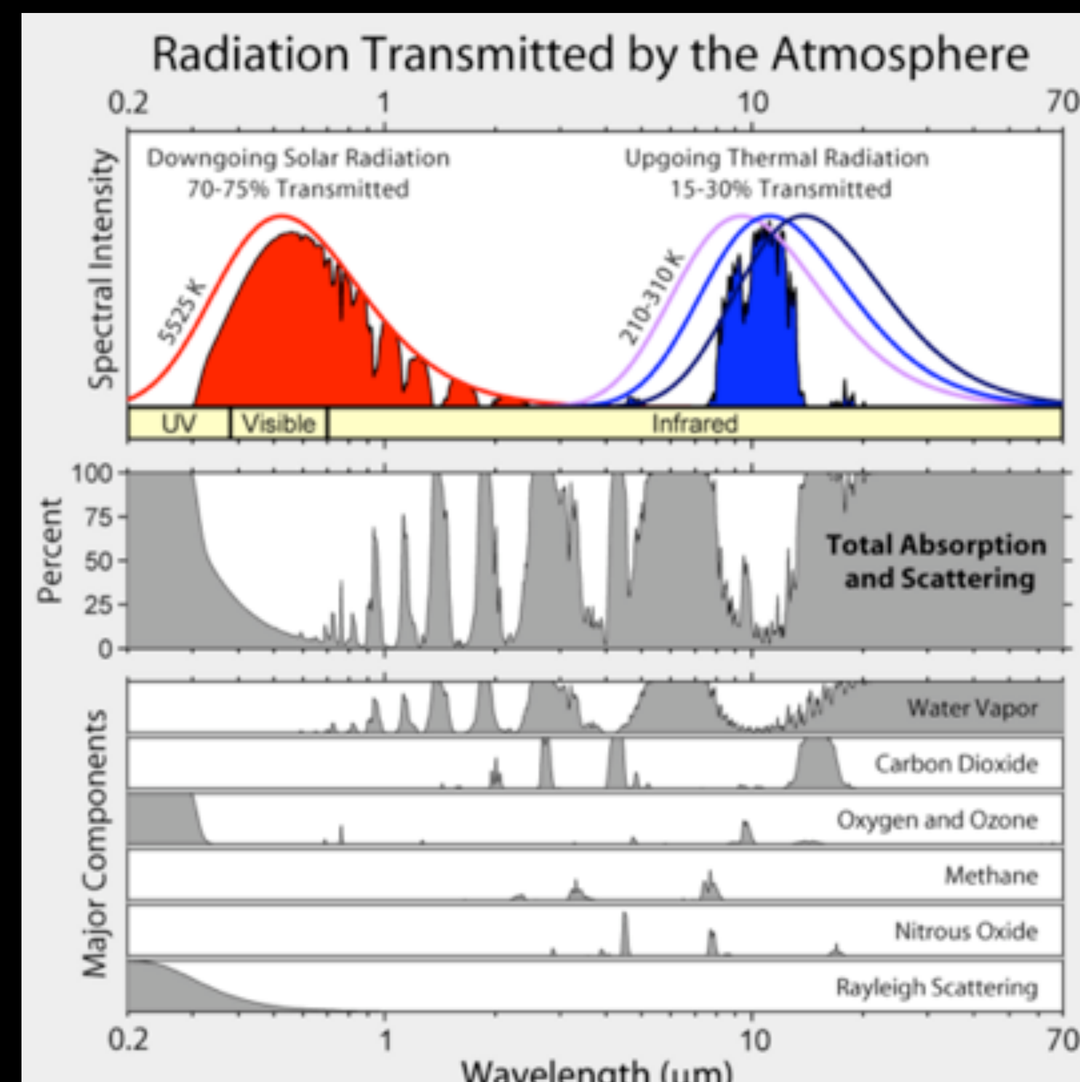




Mesosphere

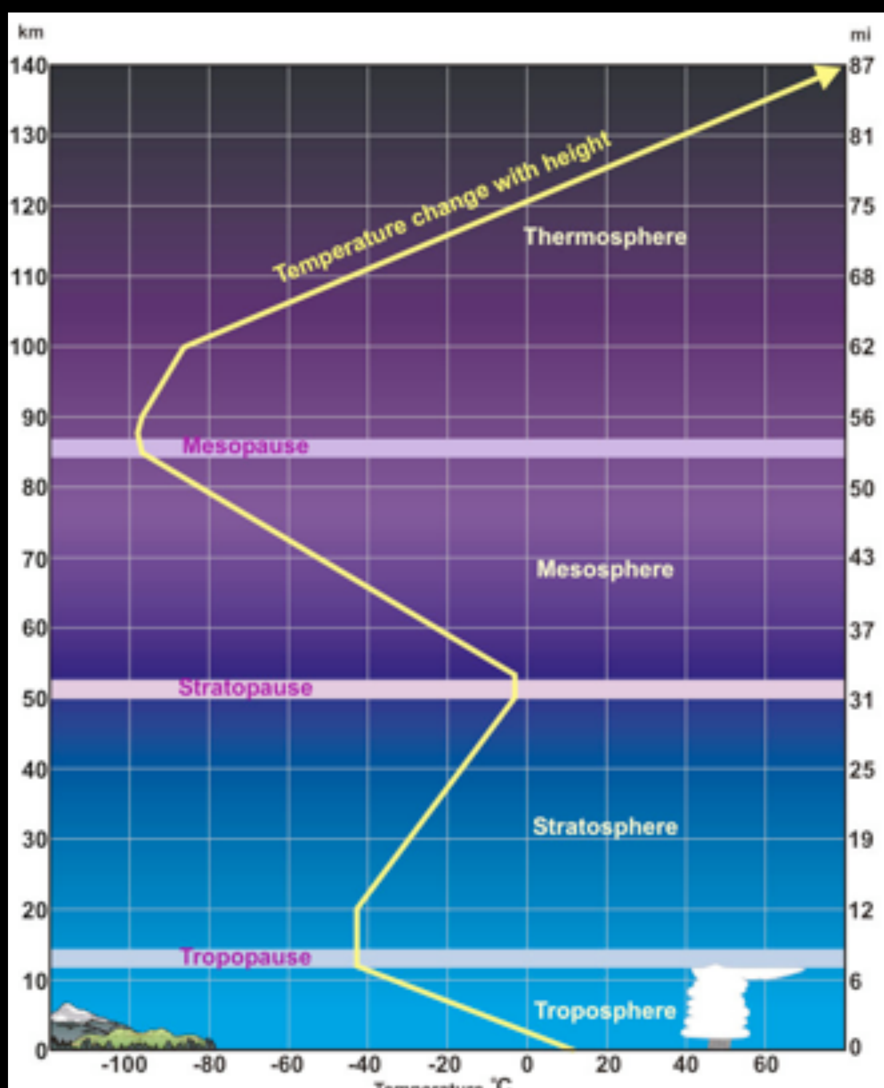
- Named from the Greek μεσό (middle) σφαιρα (sphere), located from about 50 km to 80-90 km above Earth's surface
- Composition is similar to troposphere and stratosphere (N_2, O_2, Ar, CO_2)
- Absorbs decreasing amounts of low-energy UV through ozone
- Temperature decreases with height from 0°C to -100°C

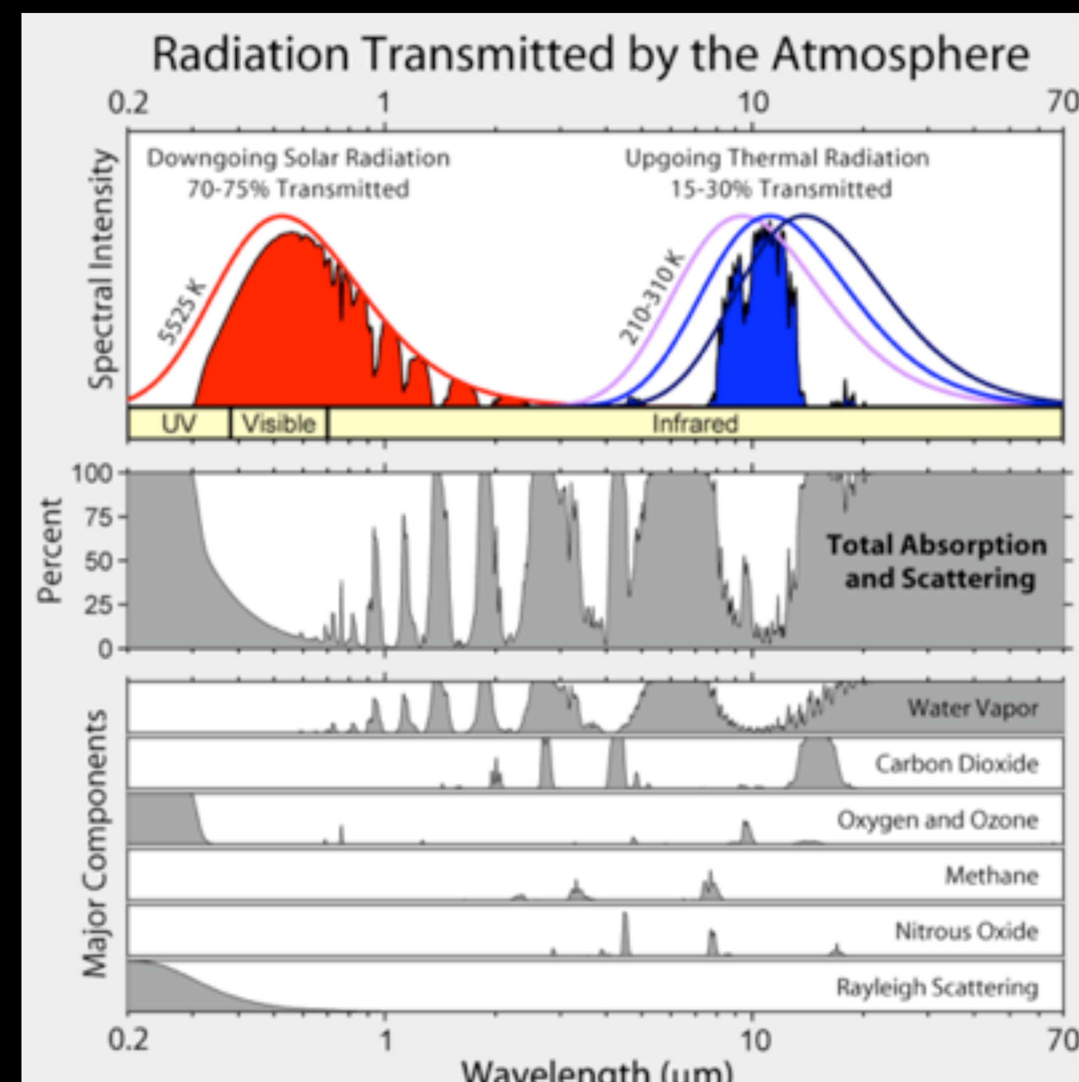




Stratosphere

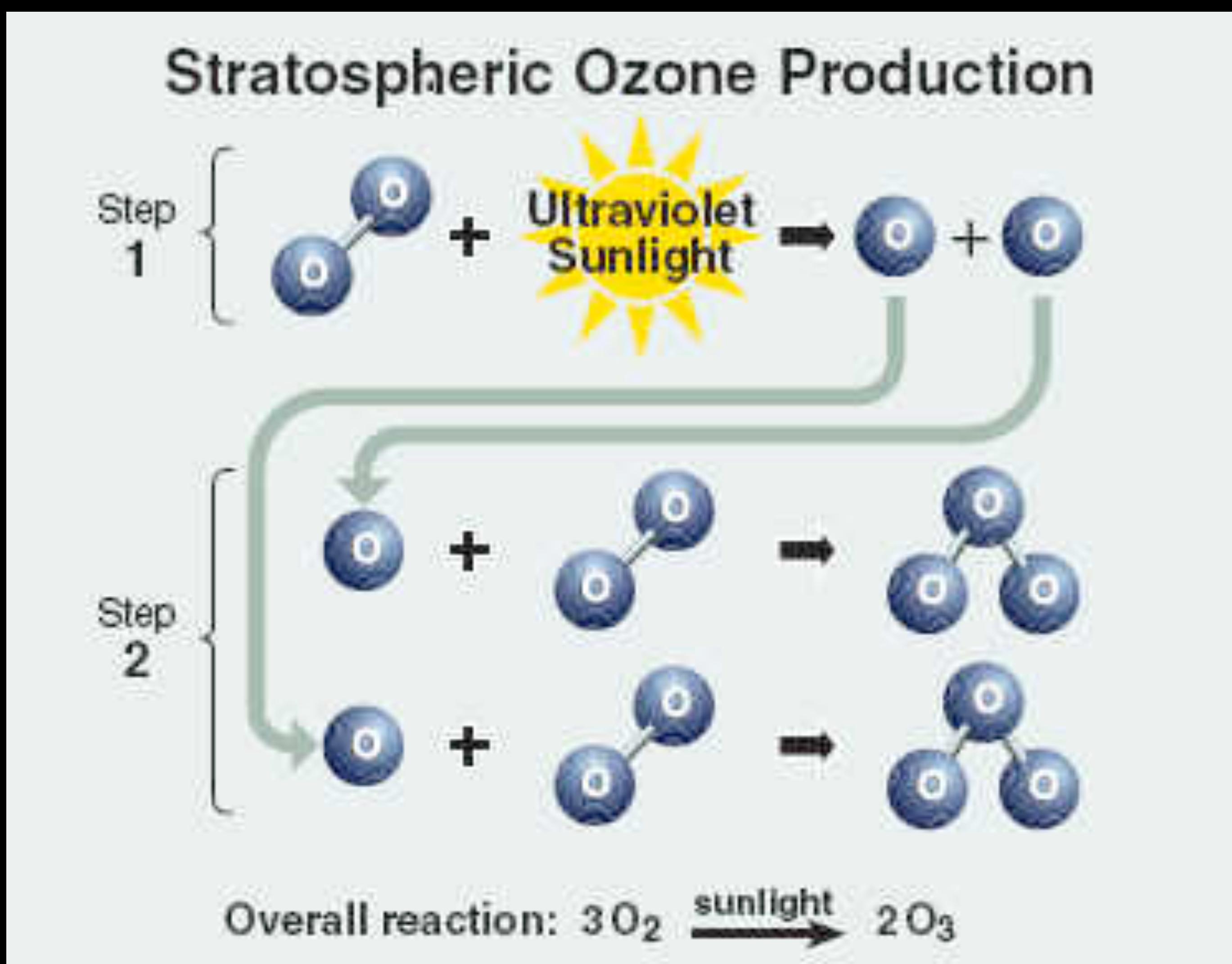
- Named from the Greek στράτος (layered) σφαιρα (sphere), located from about 10 km to 50 km above Earth's surface
- Contains most of the ozone in the atmosphere
- Ozone absorbs medium wavelength UV ($0.1 < \lambda < 0.35$ μm) - heated from above
- Temperature increases with height from -40°C to 0°C





Ozone layer

- Ozone is byproduct of photodissociation of molecular oxygen
- Ozone absorbs medium wavelength UV (important for life on Earth)



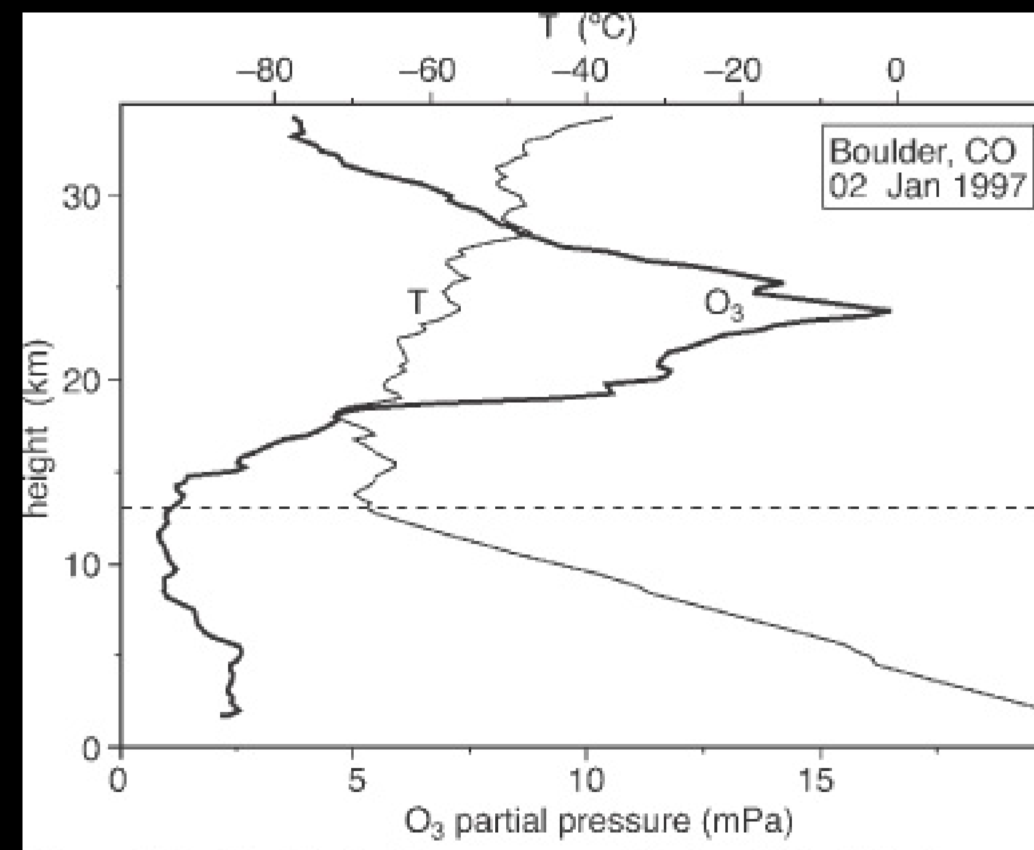
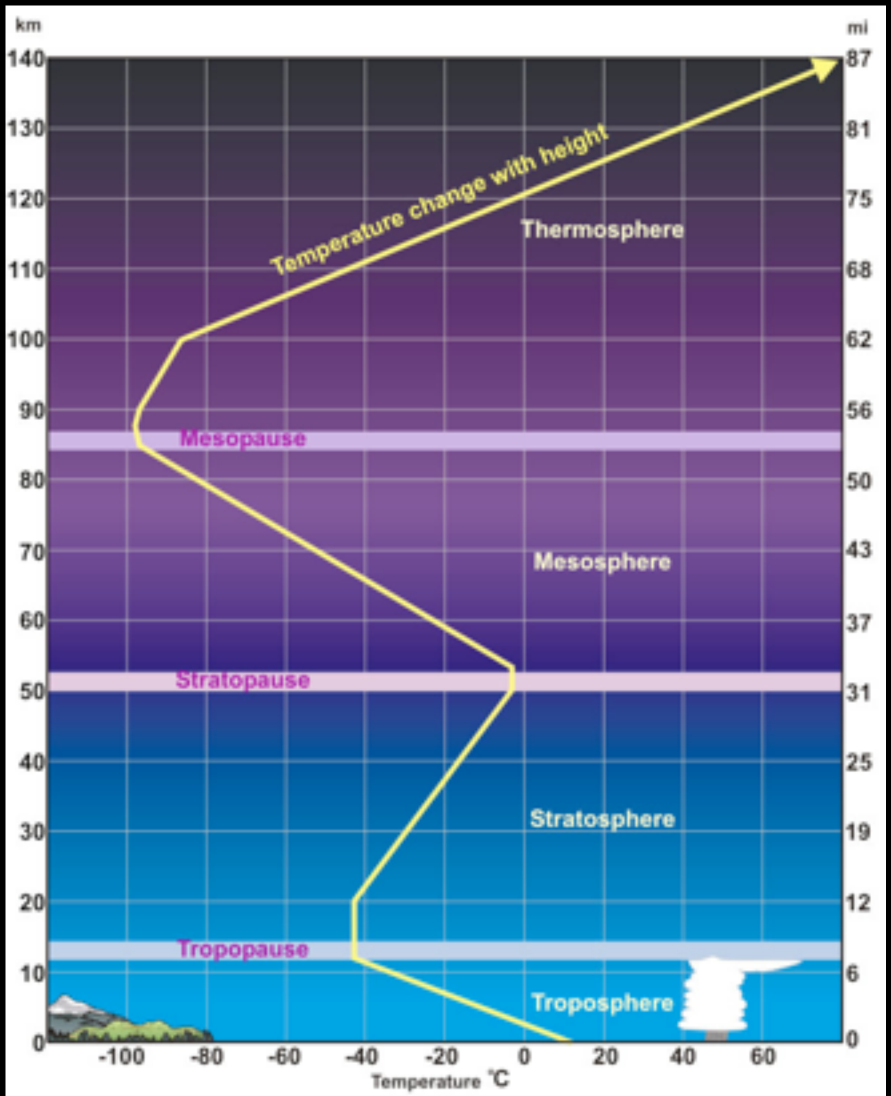
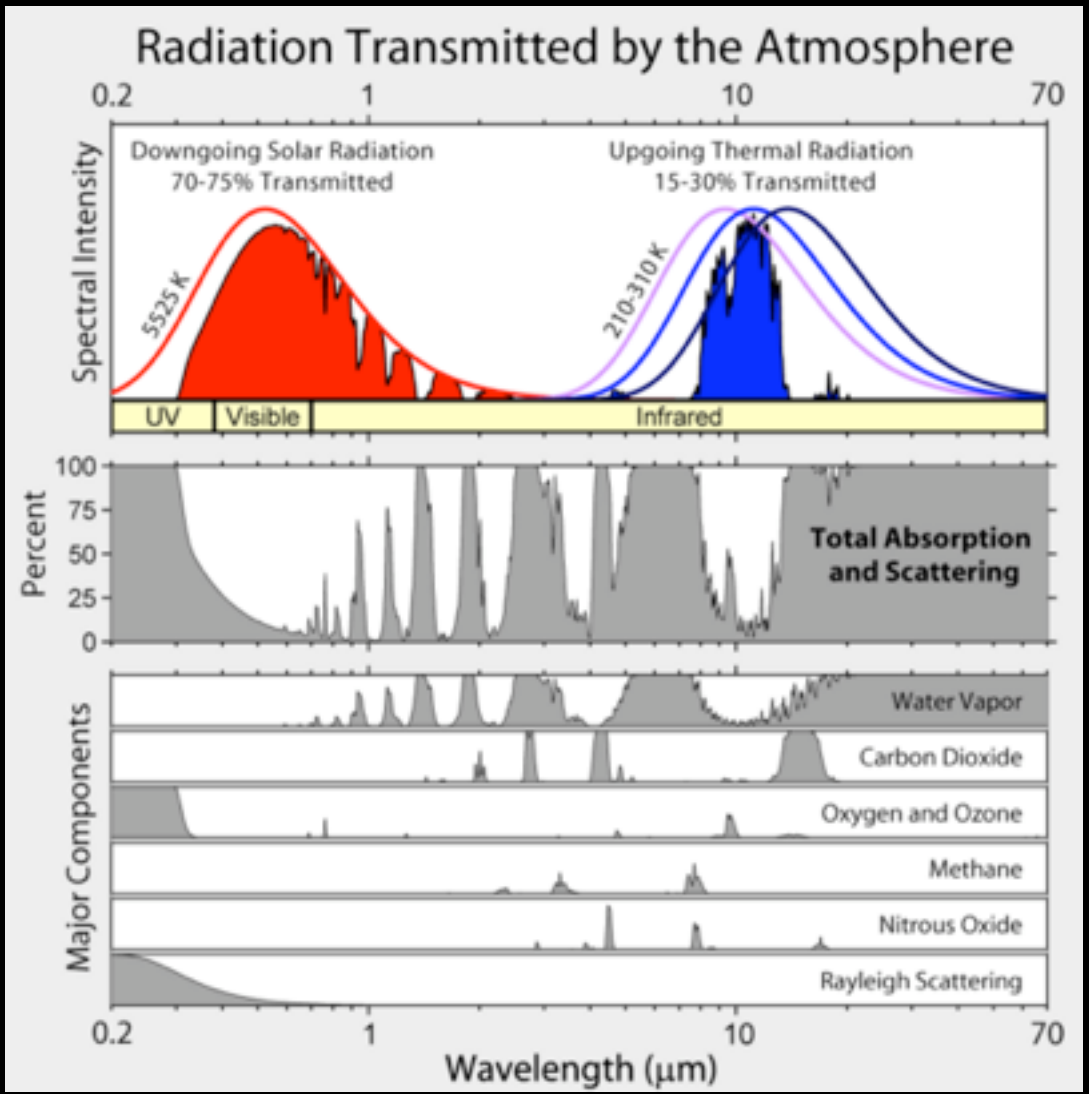


Figure 3.2: A typical winter ozone profile in middle latitudes (Boulder, CO, USA, 2 Jan 1997). The heavy curve shows the profile of ozone partial pressure (mPa), the light curve temperature (°C) plotted against altitude up to about 33 km. The dashed horizontal line shows the approximate position of the tropopause. Balloon data courtesy of NOAA Climate Monitoring and Diagnostics Laboratory.

Troposhere

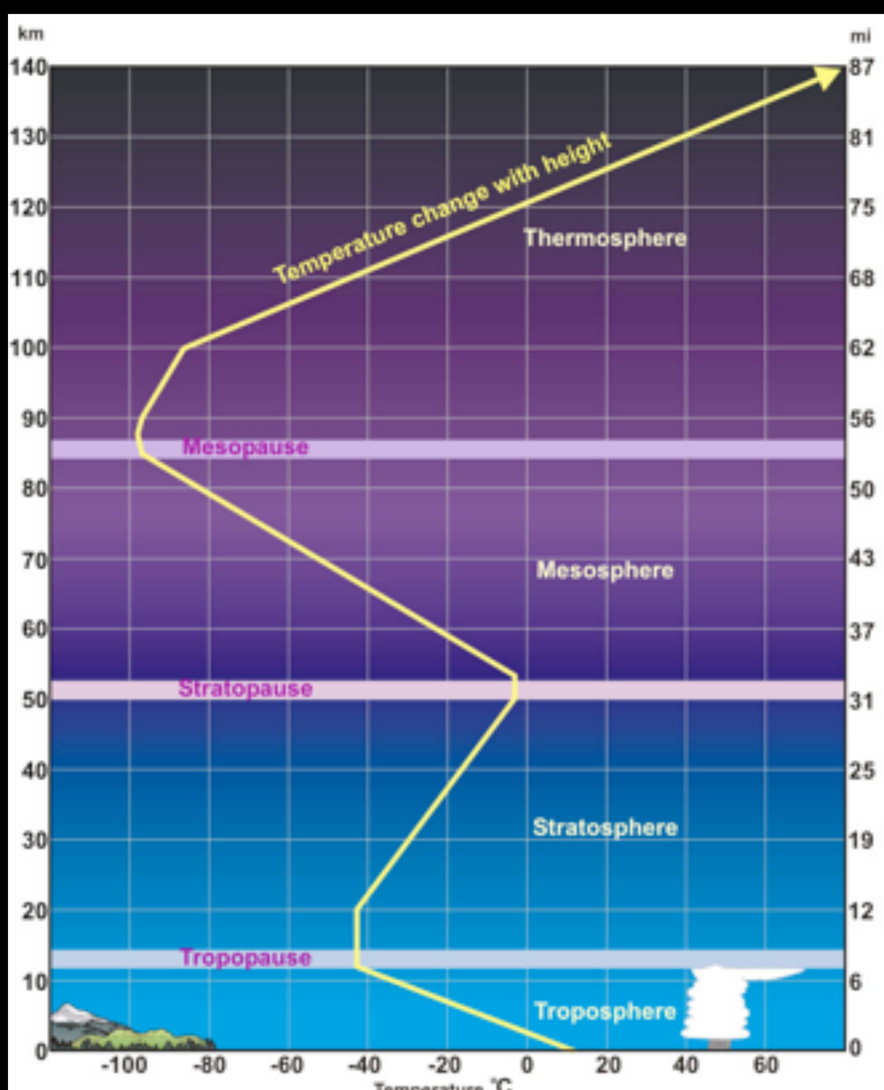
- Named from the Greek τρόπος (turn) σφαιρα (sphere), located within 10 km of the Earth's surface
- Contains 85% of the atmospheric mass
- Absorbs infrared radiation through H_2O and CO_2 - heated from below
- Temperature decreases with height from to 15°C -40°C





Troposphere

- Atmospheric layer most dynamically active
  - ➡ weather we experience is confined to troposphere
- Transport modifies temperature profiles



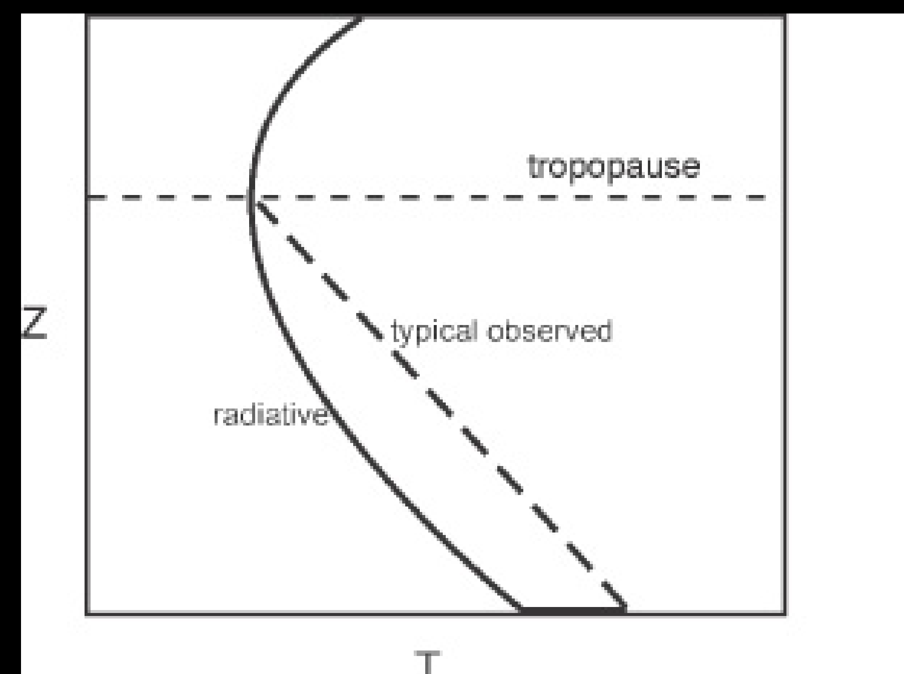


Figure 3.4: A schematized radiative equilibrium profile in the troposphere (cf. Fig. 2.11) (solid) and a schematized observed profile (dashed). Below the tropopause, the troposphere is stirred by convection and weather systems and is not in radiative balance. Above the tropopause, dynamical heat transport is less important, and the observed *T* is close to the radiative profile.

Hydrostatic balance

$$\underbrace{F_g}_{\text{Gravitational force}} + \underbrace{F_T}_{\text{Top pressure force}} + \underbrace{F_B}_{\text{Bottom pressure force}} = 0$$

$$\frac{dp}{dz} = -g\rho$$

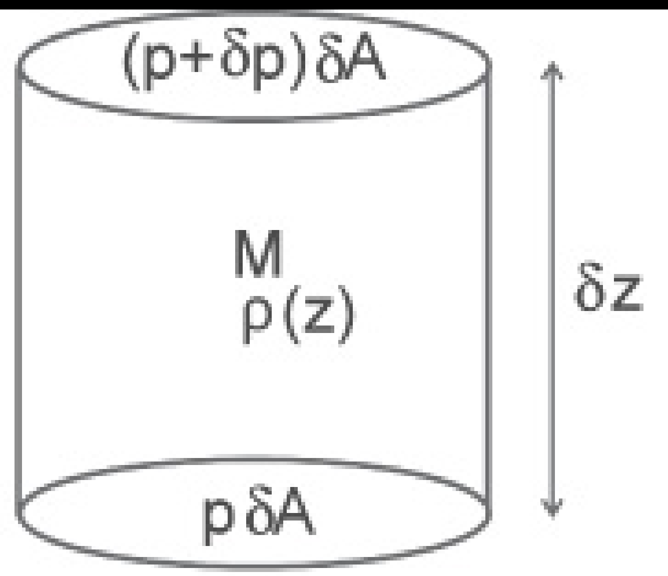


Figure 3.5: A vertical column of air of density ρ, horizontal cross-sectional area δA, height δz, and mass $M = \rho \delta A \delta z$. The pressure on the lower surface is p, the pressure on the upper surface is $p + \delta p$.

Vertical distribution of pressure

- Isothermal atmosphere

$$p(z) = p_s \exp\left(-\frac{z}{H}\right) \qquad H = \frac{RT_0}{g}$$

- Non-isothermal atmosphere

$$p(z) = p_s \exp\left(-\int_0^z \frac{dz'}{H(z')}\right) \qquad H(z) = \frac{RT(z)}{g}$$

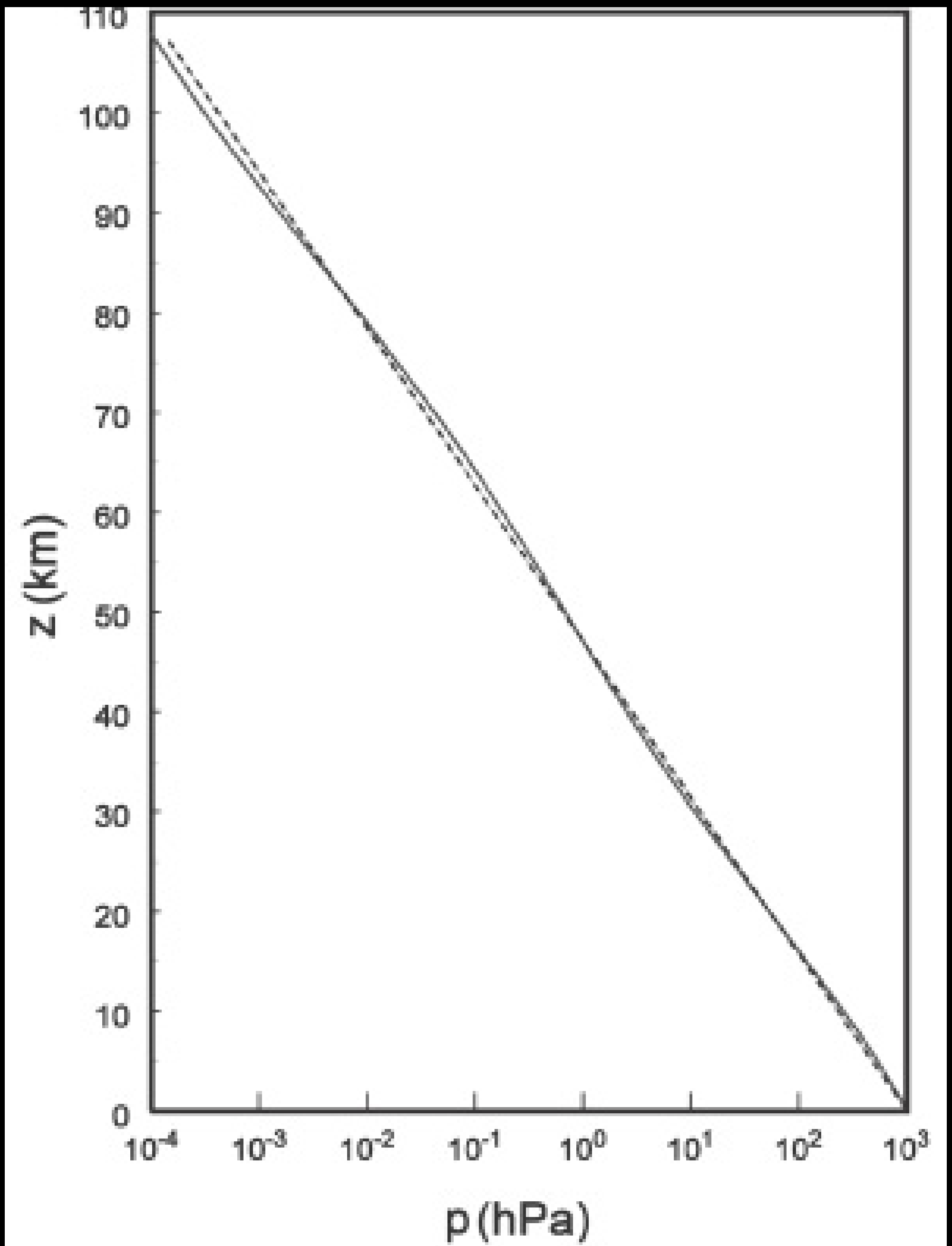


Figure 3.6: Observed profile of pressure (solid) plotted against a theoretical profile (dashed) based on Eq. 3-7 with $H = 6.8$ km.

Vertical distribution of density

- Isothermal atmosphere

$$\rho(z) = \frac{p_s}{RT_0} \exp\left(-\frac{z}{H}\right) \qquad H = \frac{RT_0}{g}$$

- Non-isothermal atmosphere

$$\rho(z) = \frac{p_s}{RT(z)} \exp\left(-\int_0^z \frac{dz'}{H(z')}\right) \qquad H(z) = \frac{RT(z)}{g}$$