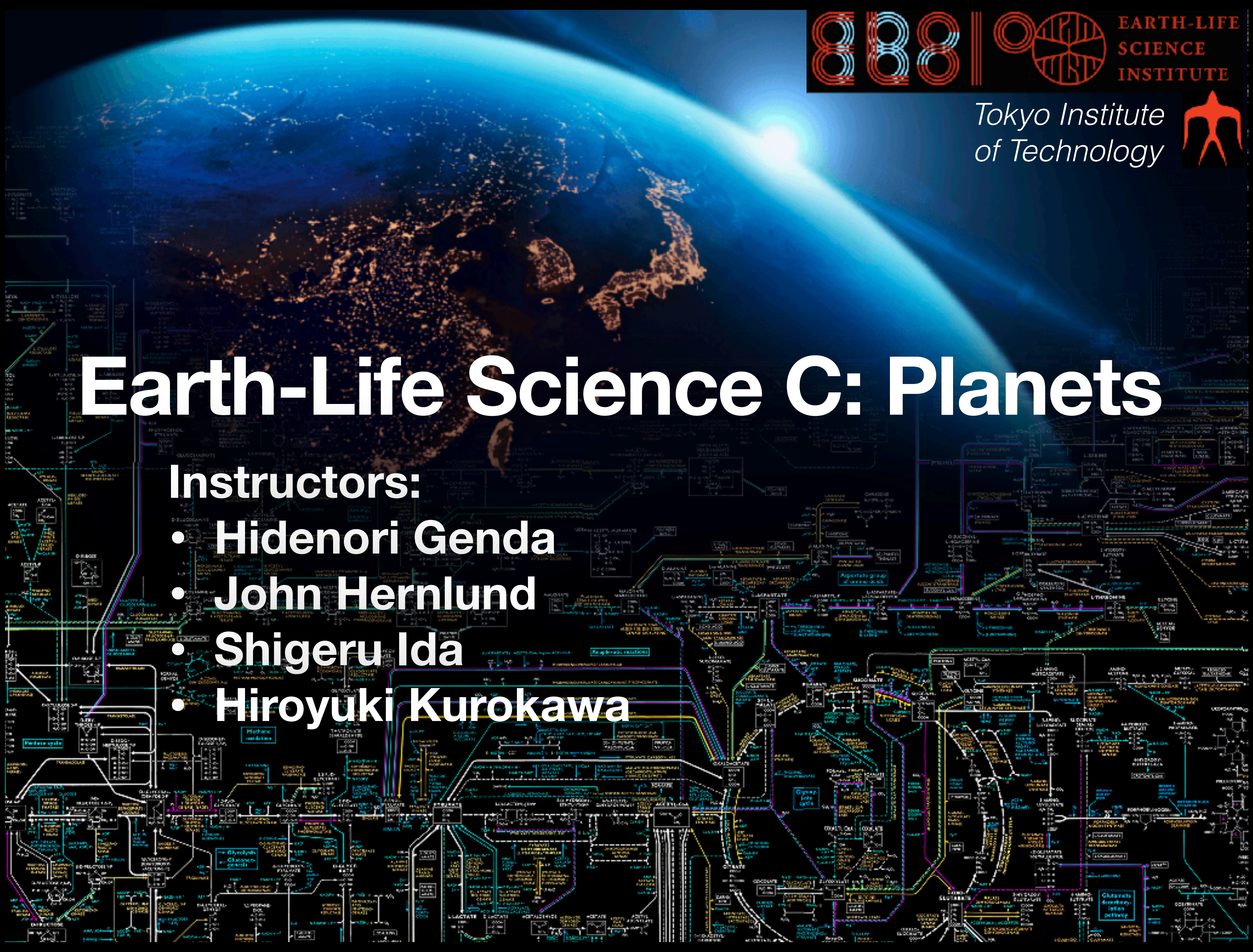




Earth-Life Science C: Planets

Instructors:

- Hidenori Genda
- John Hernlund
- Shigeru Ida
- Hiroyuki Kurokawa



Report assignment of the previous lecture

Summarize your answers into a short report and submit it by the beginning of the next lecture (either directly, to my post-box, or by e-mail to hiro.kurokawa@elsi.jp).

1. Thermal conduction determines the temperature profile in the boundary layer. Using the physical quantities given below, estimate the temperature gradient in the upper boundary layer of Earth's interior (the top ~ 100 km). Answer with one significant digit.

$$F_{\text{int}} = 0.09 \text{ W} \cdot \text{m}^{-2}, \kappa \simeq 1 \times 10^{-6} \text{ m}^2 \cdot \text{s}^{-1}, \rho \simeq 3 \times 10^3 \text{ kg} \cdot \text{m}^{-3}, c_p \simeq 1 \times 10^3 \text{ J} \cdot \text{kg}^{-1} \cdot \text{K}^{-1}$$

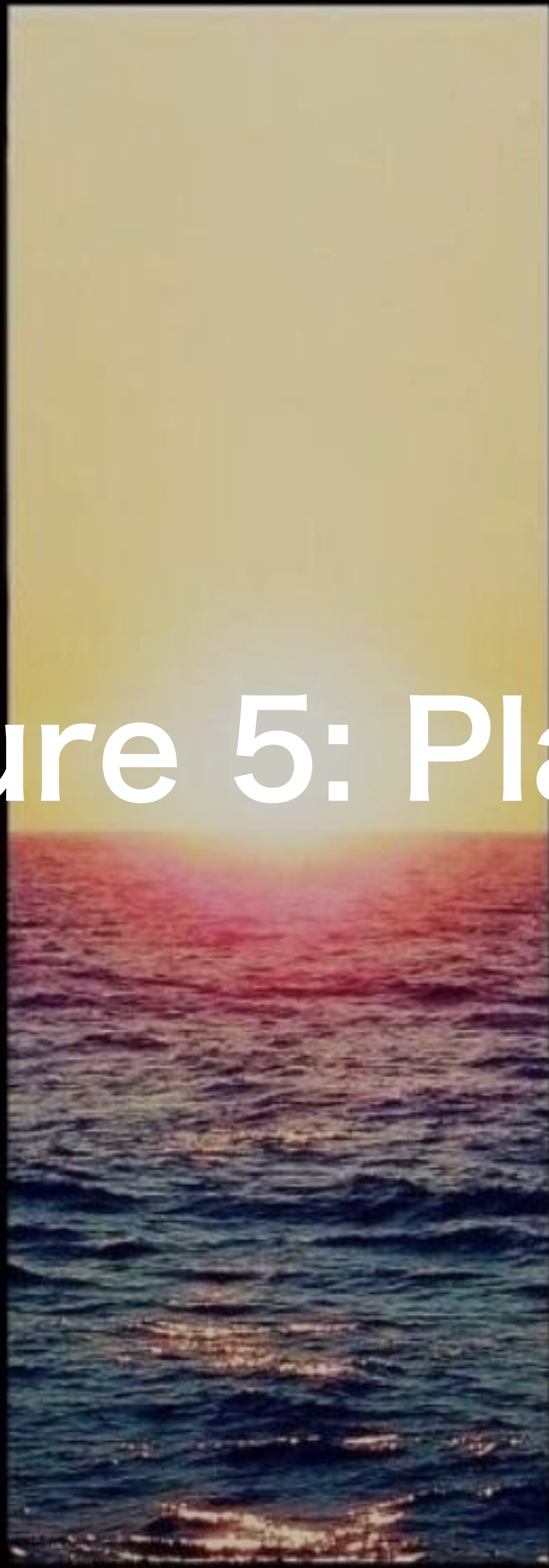
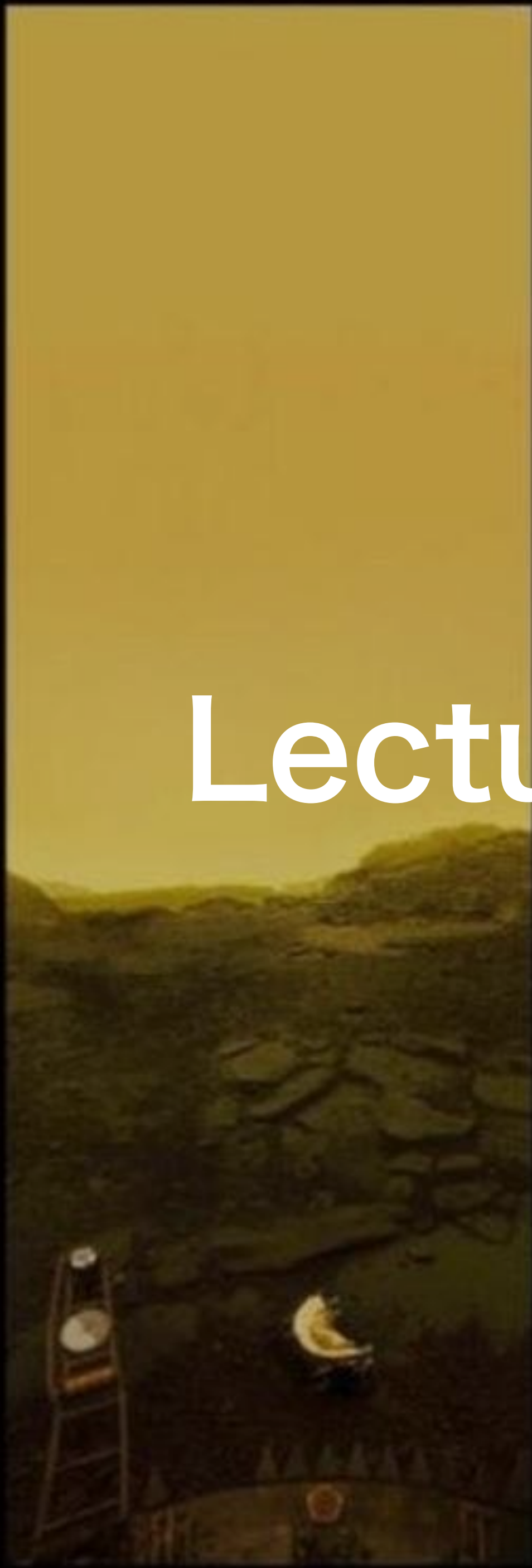
$$\therefore \left(\frac{dT}{dz} \right)_{\text{cond}} = - \frac{F_{\text{int}}}{k_{\text{cond}}} = - \frac{F_{\text{int}}}{\rho C_p \kappa} \simeq - 30 \text{ K} \cdot \text{km}^{-1}$$

2. Let's assume that you are a hot-spring (*onsen*) enthusiast and want to dig for a hot spring of your own. Using the result of Q1, discuss how deep you need to dig a hole in the ground.

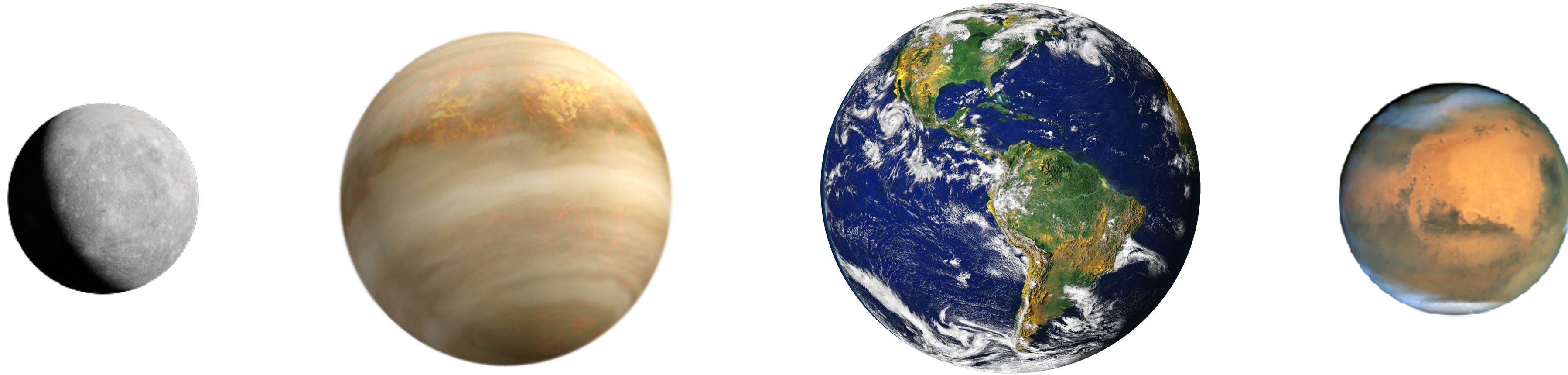
Assuming that I want to have 45°C hot water and the surface temperature is 15°C , I need to dig a hole of 1 km depth.

- Hernlund**
 - Lecture 1: The present-day Earth (Tuesday, 4 October)
 - Lecture 2: Earth's history (Friday, 7 October)
 - Lecture 3: Exploration of the Solar System (Tuesday, 11 October)
- Kurokawa**
 - Lecture 4: Planetary structure and equations (Friday, 14 October)
 - Lecture 5: Planetary atmospheres (Tuesday, 18 October)
 - Lecture 6: Climate evolution, volatile cycling, and biogeochemical cycling (Friday, 21 October)
- Ida**
 - Lecture 7: Planet formation (~~Tuesday, 1 November~~ **Friday, 28 October**)
 - Lecture 8: Satellite formation (Friday, 4 November)
 - Lecture 9: Origins of organic materials (Tuesday, 8 November)
- Genda**
 - Lecture 10: Water delivery to Earth (Friday, 11 November)
 - Lecture 11: Stellar evolution (Tuesday, 15 November)
 - Lecture 12: Exoplanet observations (Friday, 18 November)
- All**
 - Lecture 13: Summary and future prospects (Tuesday, 22 November)

Lecture 5: Planetary atmospheres



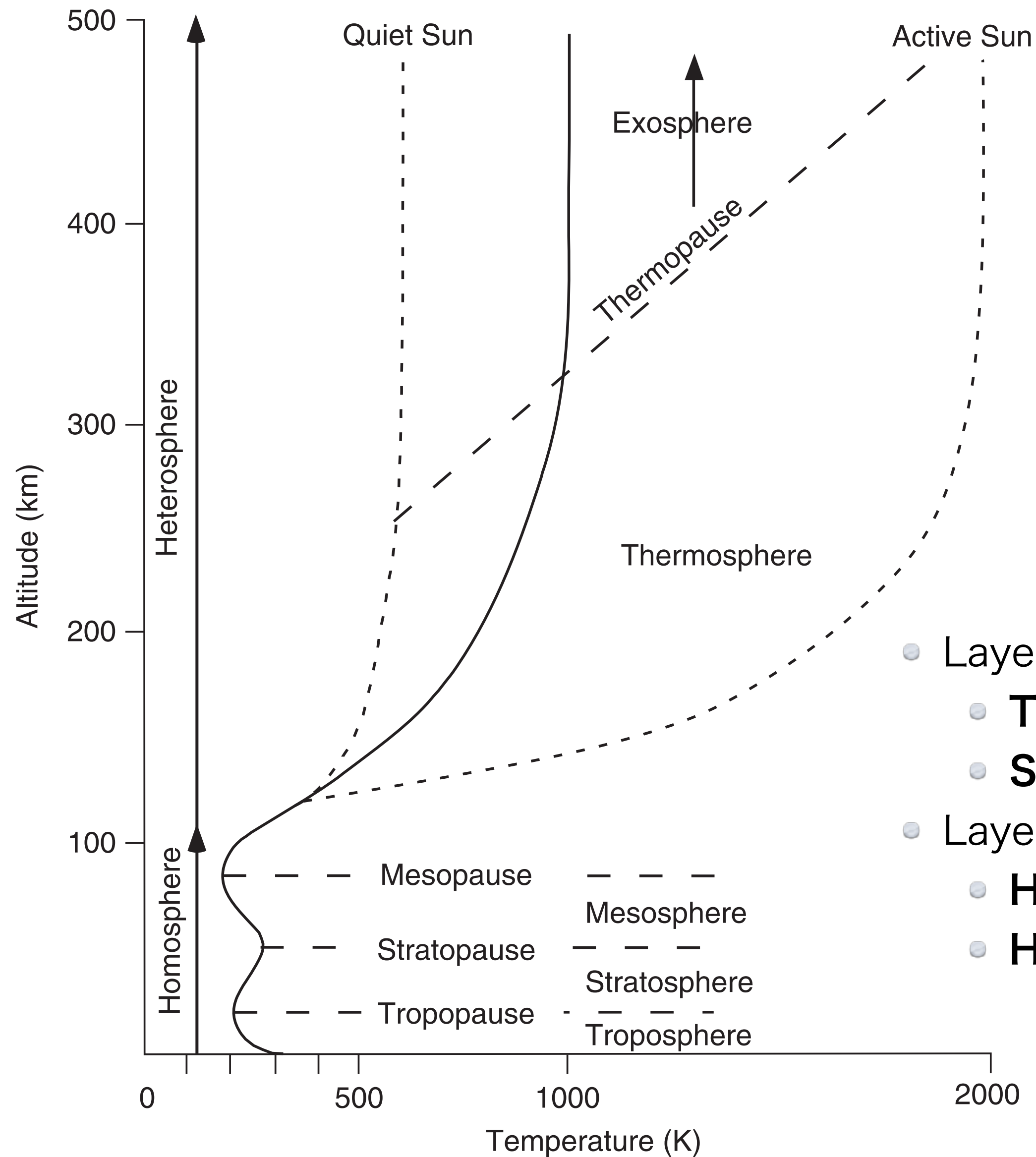
Atmospheres of terrestrial planets



	Mercury	Venus	Earth	Mars
Orbital radius [au]	0.4	0.7	1	1.5
Pressure [bar]	—	90	1	0.006
Composition	—	CO ₂ (>95%)	N ₂ , O ₂	CO ₂ (>95%)
Temperature [K]	440	740	288	210
Water mass [Earth=1]	—	10 ⁻⁵ (vapor)	1 (liquid)	10 ⁻³ (ice)

Note: 10⁵ Pa = 1 bar ≈ 1 atm

Vertical structure of Earth's atmosphere



- Layers defined with the temperature profile:
 - **Troposphere** (convective)
 - **Stratosphere, mesosphere, thermosphere** (not convective)
- Layers defined with the chemical composition
 - **Homosphere** (homogeneous except for water vapor, ozone, etc.)
 - **Heterosphere** (heterogeneous)

Catling & Kasting (2017)
Atmospheric Evolution on Inhabited and Lifeless worlds

Rarefied upper atmosphere



Pressure & density profiles

Because Earth's atmosphere is thin compared to Earth's radius, we can approximate the hydrostatic equation as,

$$\frac{dp}{dr} = -\frac{GM}{r^2}\rho \sim \rho g \quad \text{--- (1) (} g \text{ is the gravitational acceleration at the surface).}$$

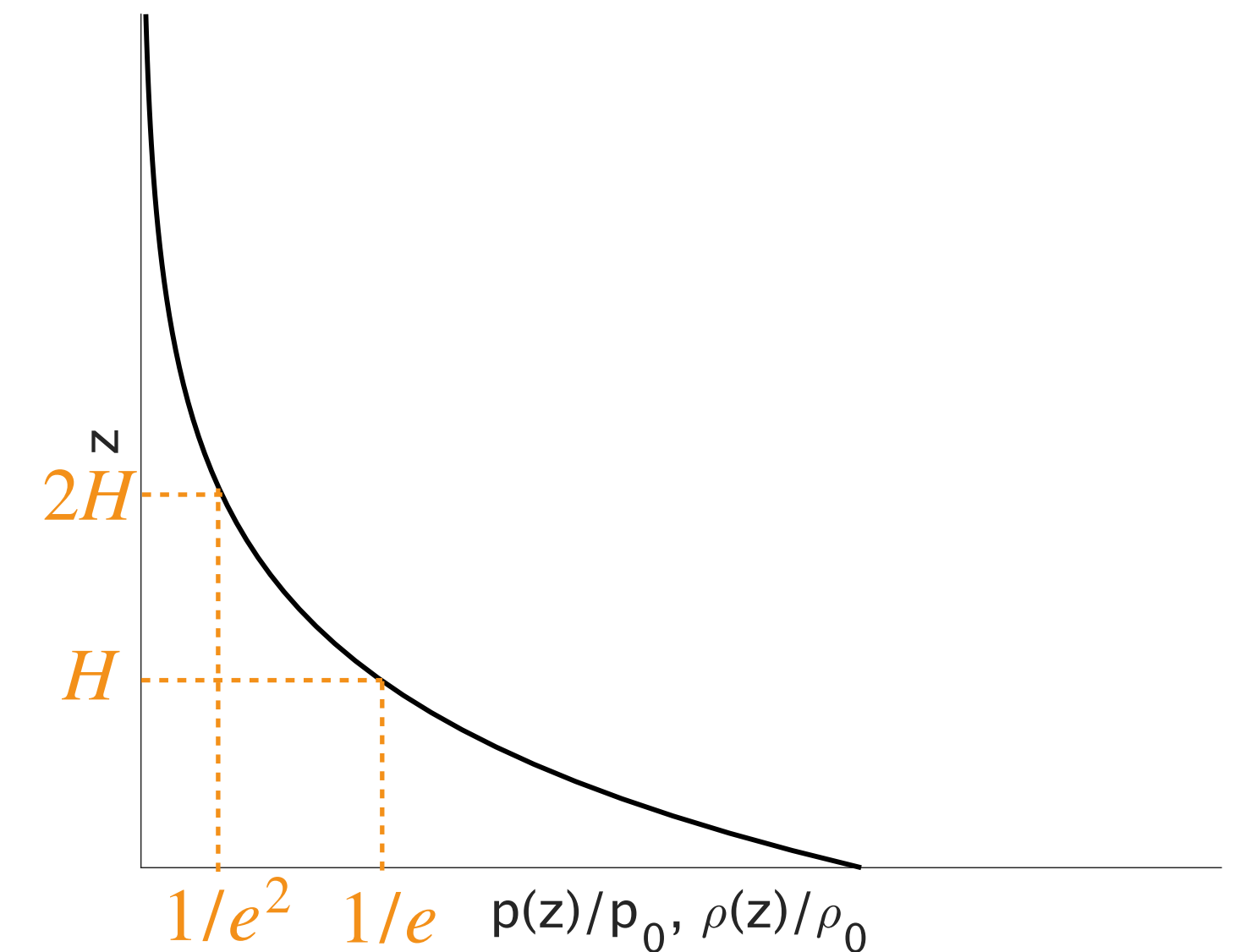
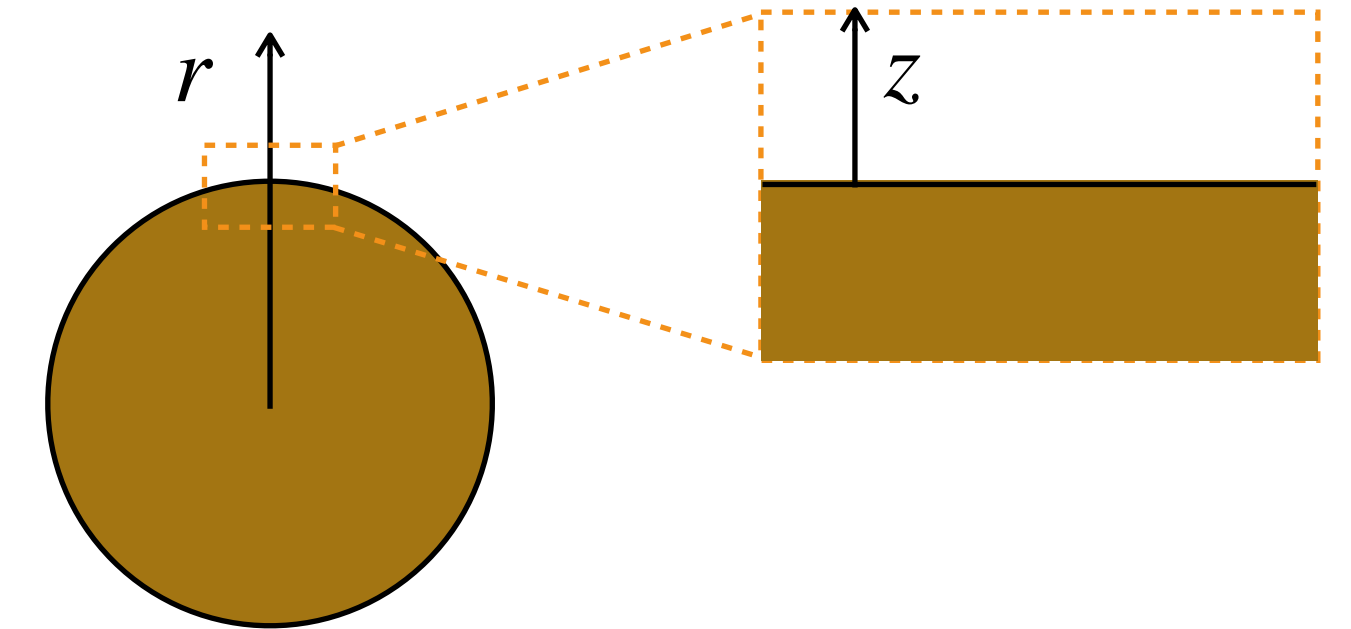
Assuming isothermal and the ideal gas law, $p = \frac{\rho k_B T}{\bar{m}}$ --- (2), we obtain,

$$\frac{dp}{dz} = \frac{\bar{m}g}{k_B T} p \quad \text{--- (3), where } z \text{ is the height from the surface.}$$

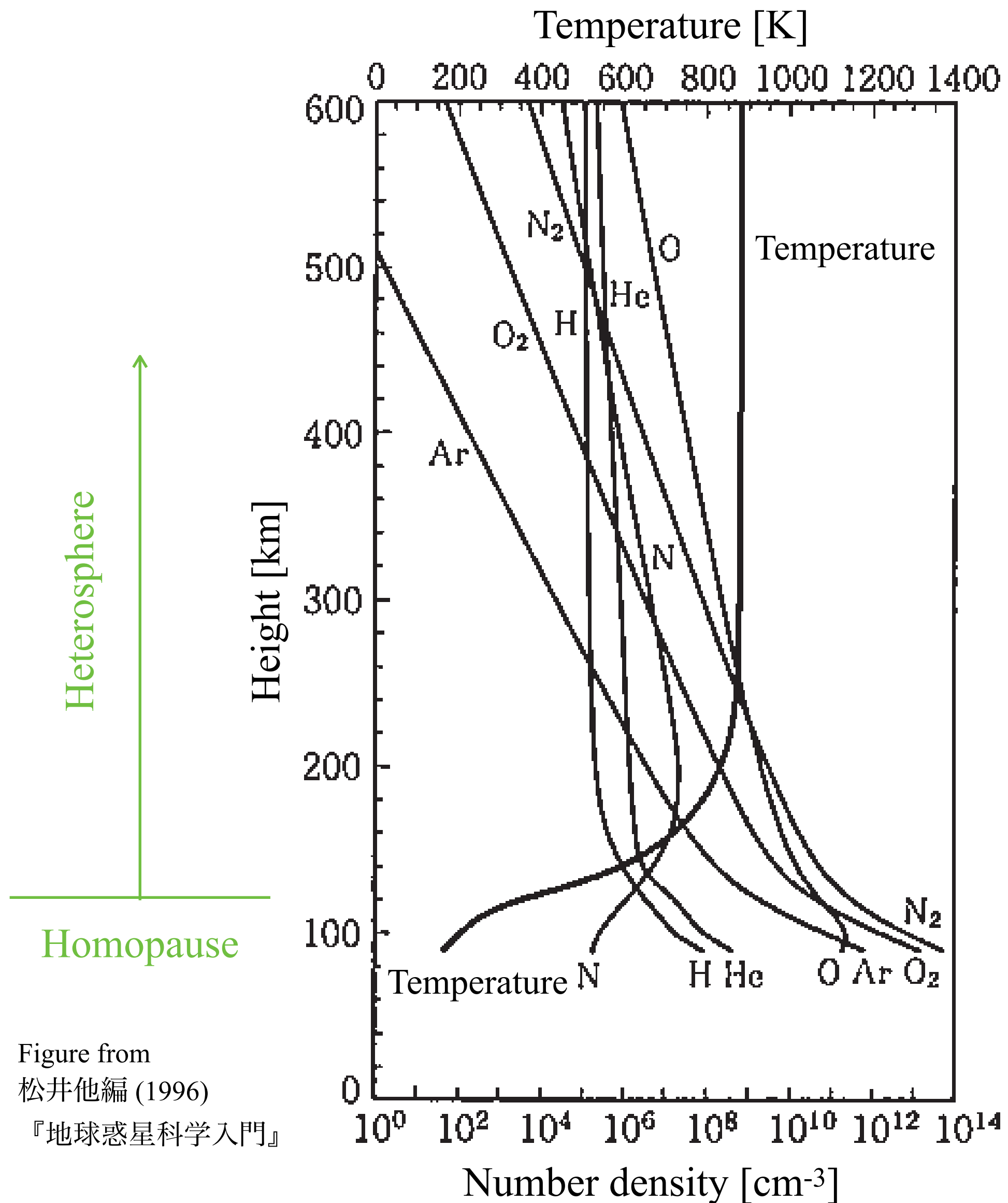
$$\therefore \frac{1}{p} dp = \frac{1}{H} dz \quad \text{--- (4), where } H \equiv \frac{k_B T}{\bar{m}g} \quad \text{--- (5) is called } \textit{the scale height}.$$

Finally, by integrating Eq. 4, we obtain $p(z) = p_0 \exp\left(-\frac{z}{H}\right)$ --- (6).

→ Both p, ρ decrease by $1/e$ every distance H !



Compositional profile in the heterosphere

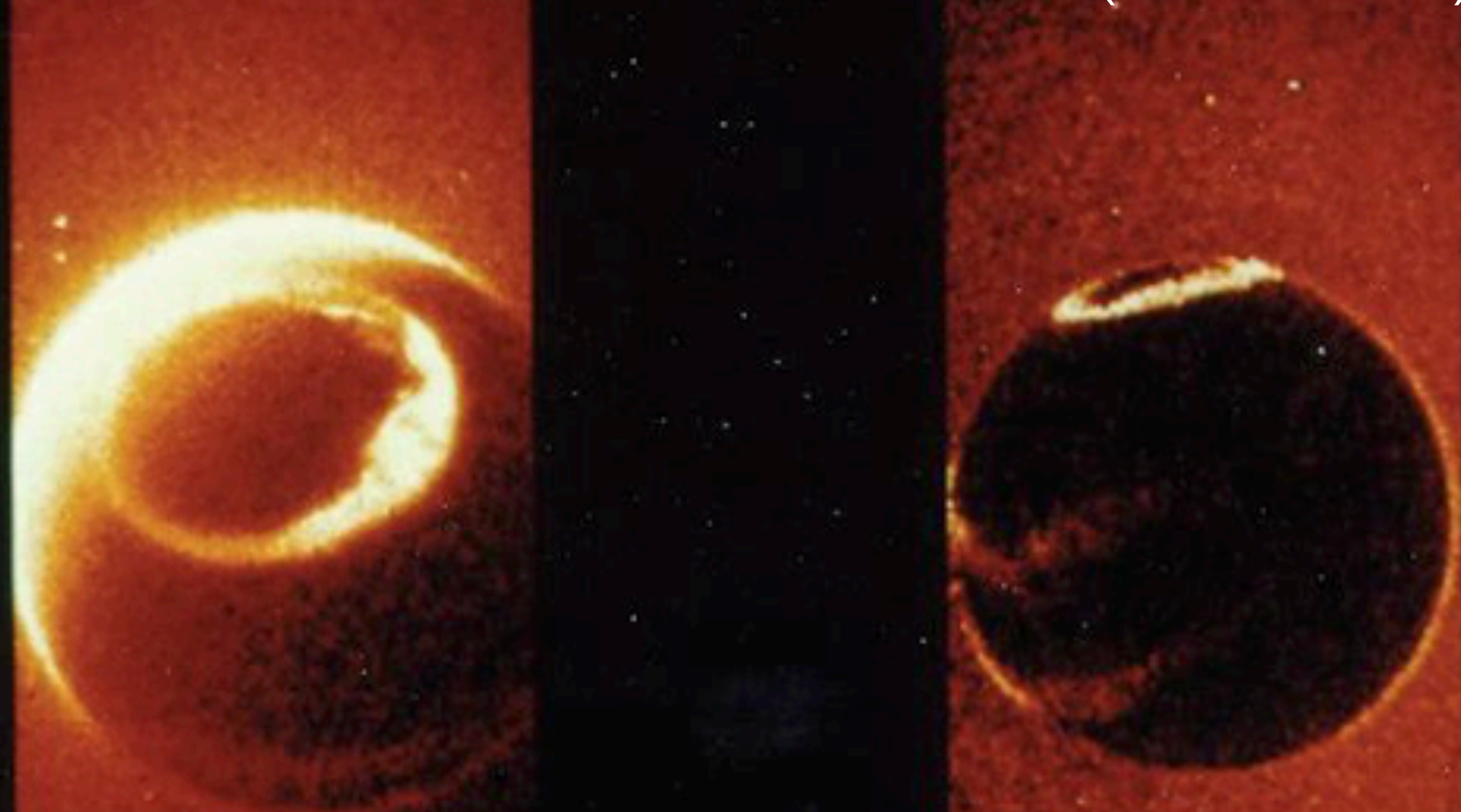


- Lower density → Longer mean free path
- In the heterosphere, lower mass molecules have larger scale heights.
 - The number density of gas species i is given by, $n_i \propto \exp(-z/H_i) = \exp(-zm_i g/k_B T)$ where H_i , m_i are the scale height and the molecular mass
- Ultimately, the molecules become collisionless → exosphere

Figure from
松井他編 (1996)
『地球惑星科学入門』

Atmospheric escape

Earth observed in vacuum-ultraviolet (100-200 nm)

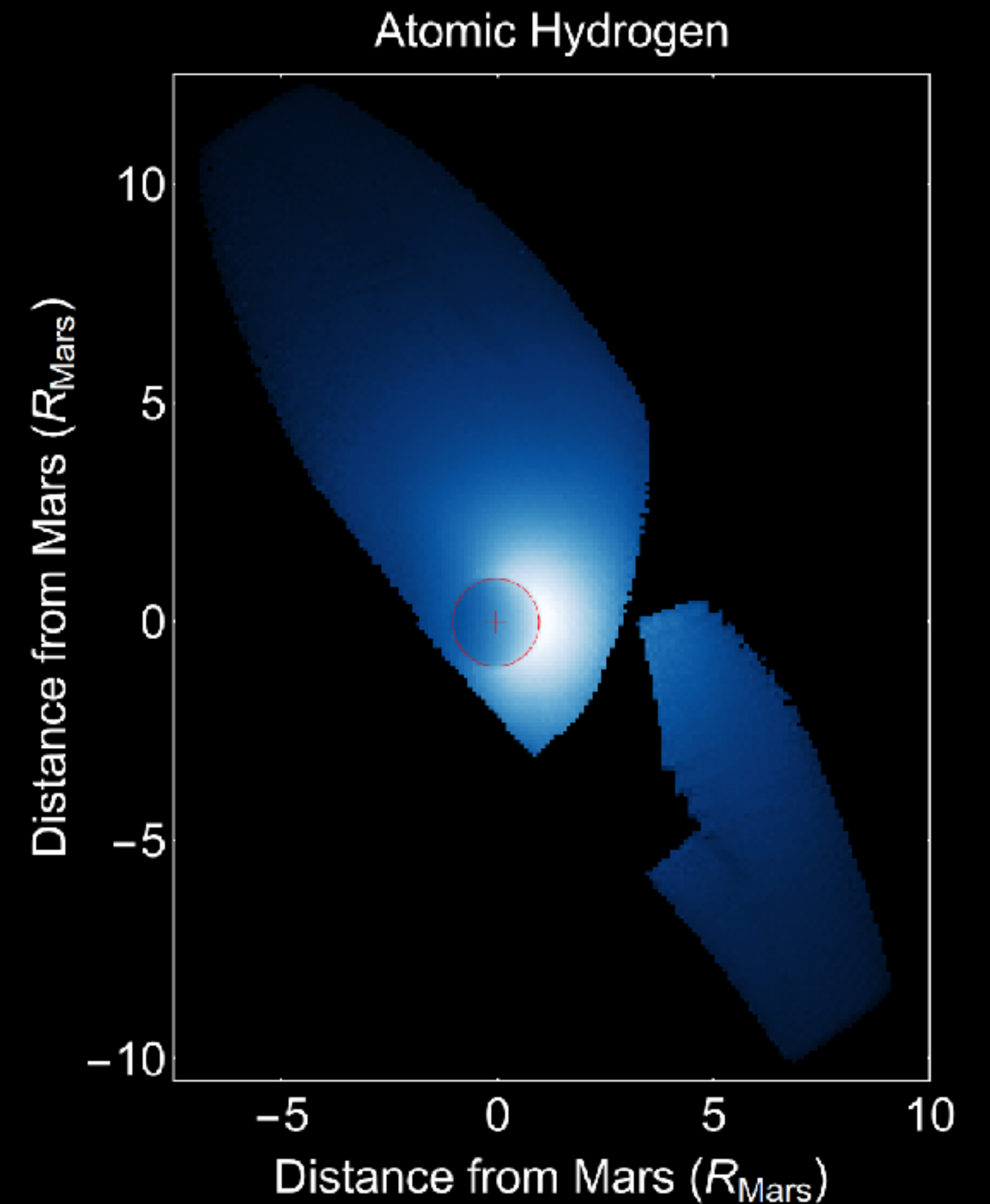
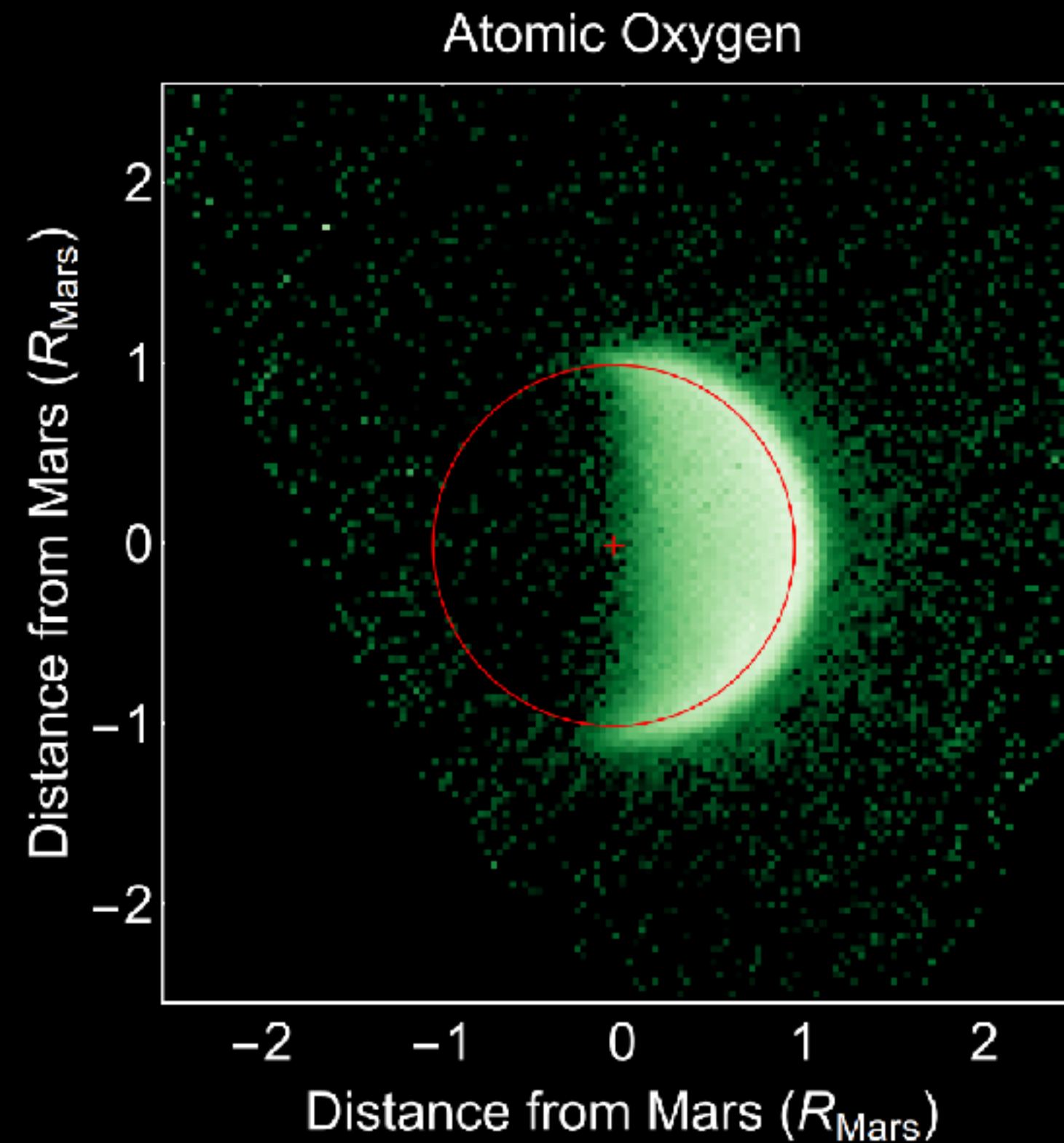
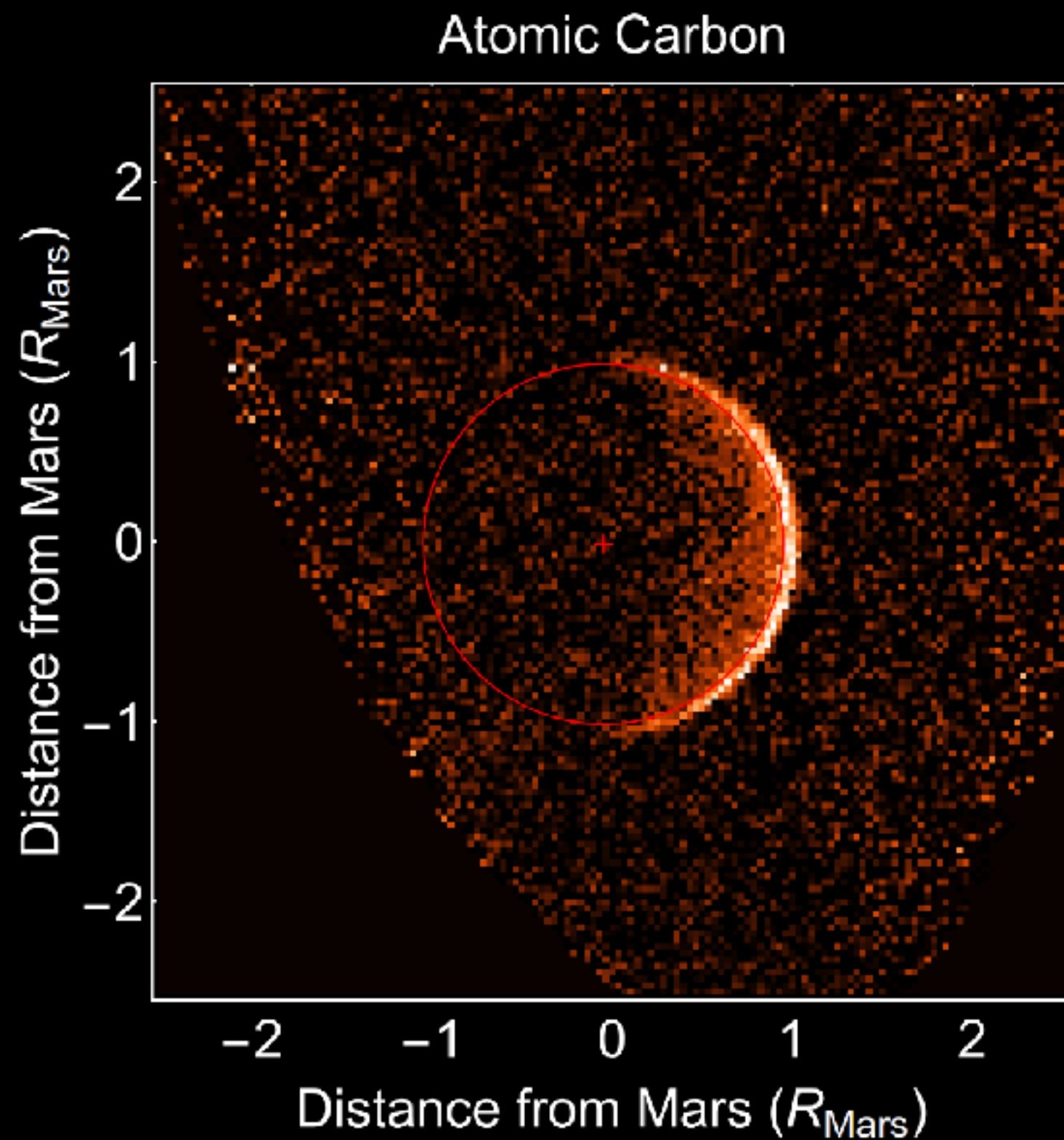


- Hydrogen surrounding Earth scatters the solar radiation
- The hydrogen originate from Earth's atmosphere (ultimately water vapor from the oceans)

Rairden et al. (1986)

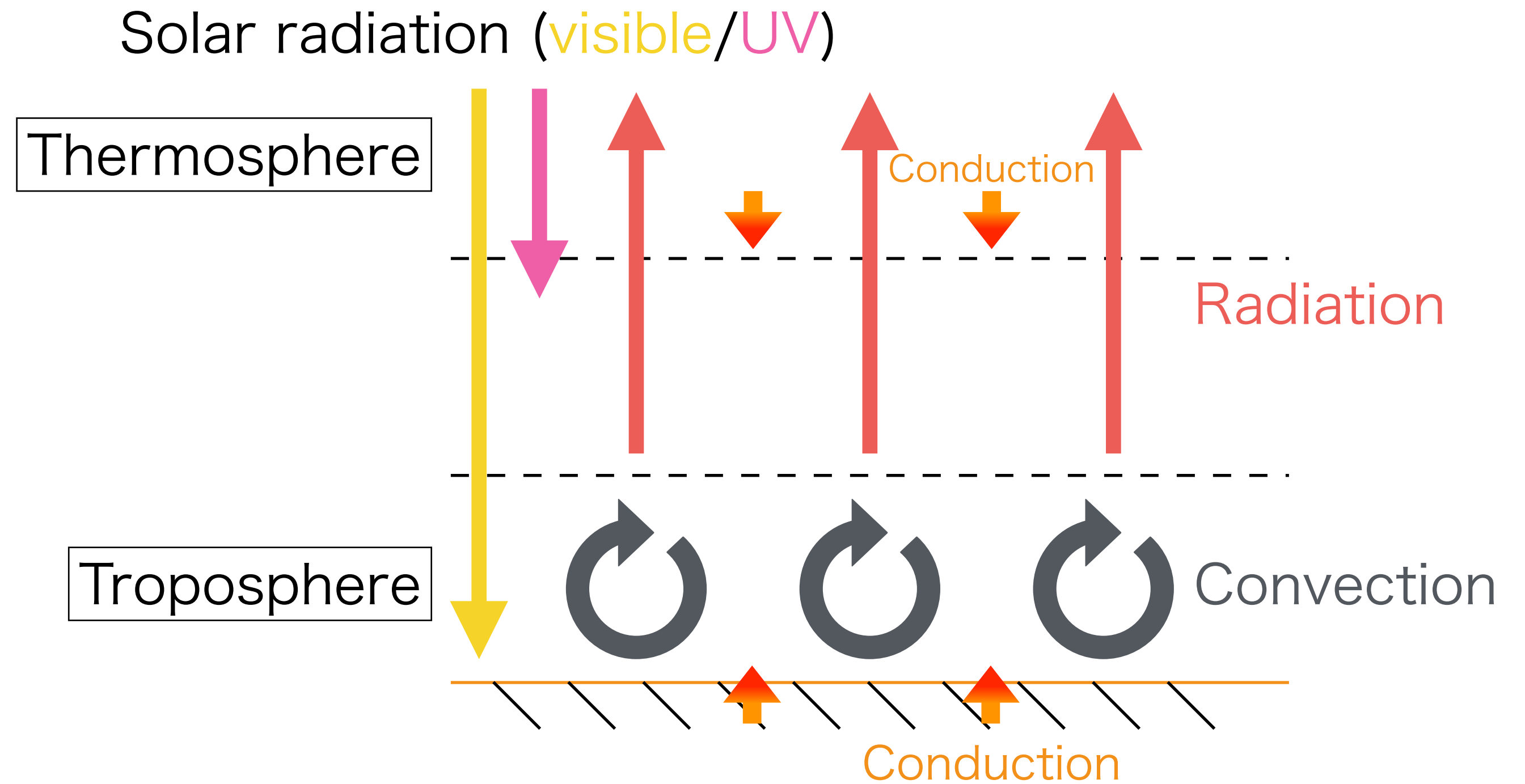
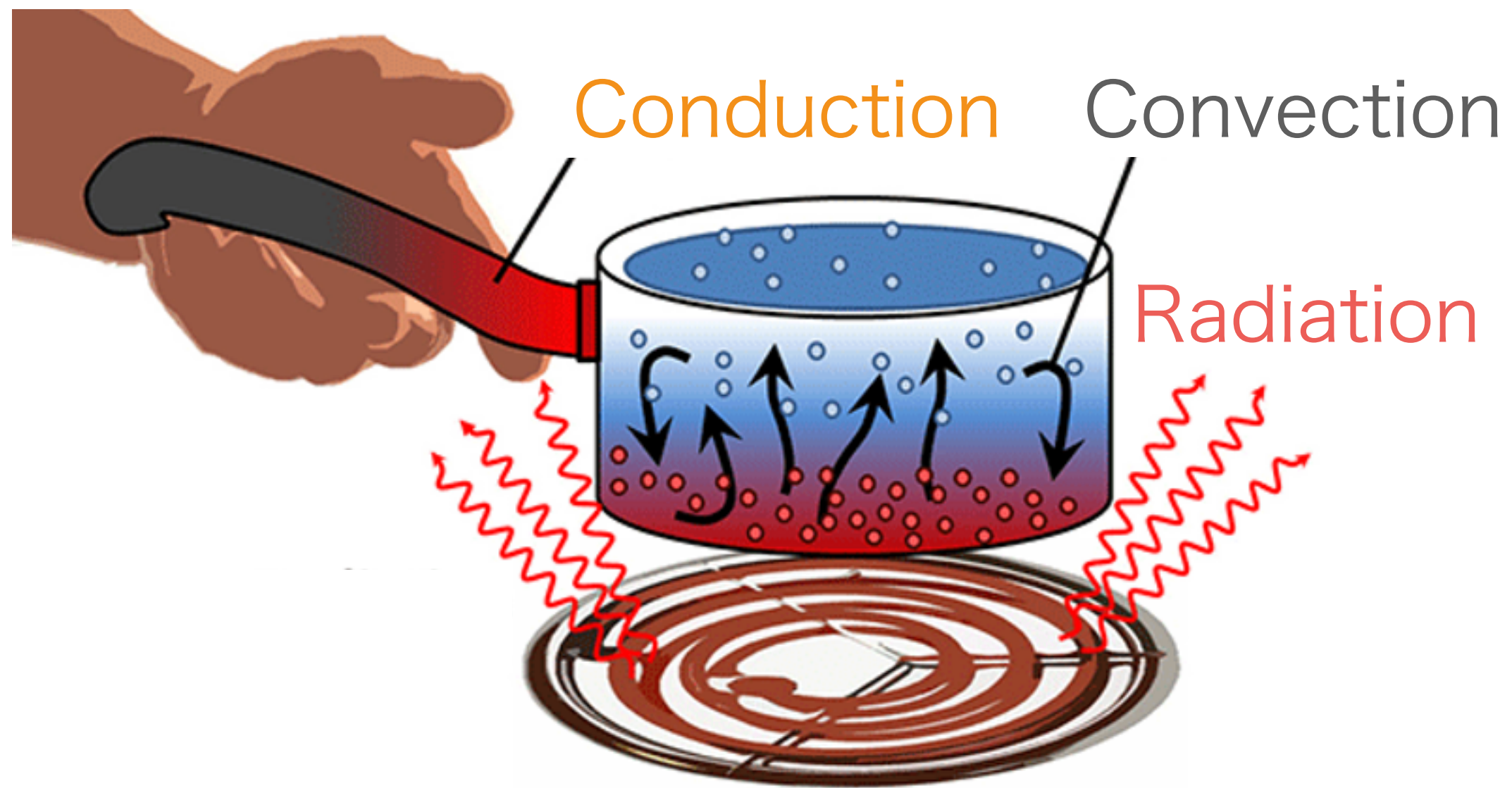
Escaping heavy atoms from Mars

Mars observed in UV with the MAVEN spacecraft (credit: Univ. Colorado, NASA)



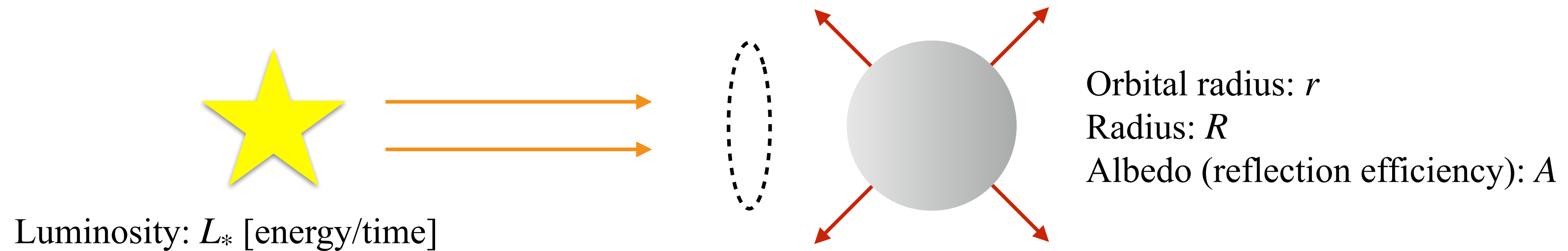
The absence of magnetic field → Atmospheric escape due to the solar wind

Energy transfer in the atmosphere



- A troposphere develops in the lower atmosphere
- Above the tropopause, energy is transferred mainly by radiation

Equilibrium Temperature T_{eq}



Let us consider the energy balance of the planet.

$$\text{Heating rate: } \dot{Q}_* = \pi R^2 \cdot \frac{L_*}{4\pi r^2} \cdot (1 - A) \quad \text{--- (1),}$$

$$\text{Cooling rate } \dot{Q}_p = 4\pi R^2 \cdot \sigma_{\text{SB}} T_{\text{eq}}^4(r) \quad \text{--- (2), where } \sigma_{\text{SB}} = 5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4} \text{ (the Stefan–Boltzmann constant)}$$

$$\text{From (1) = (2), we obtain } T_{\text{eq}} \simeq 255 \left(\frac{1 - A}{1 - A_{\oplus}} \right)^{\frac{1}{4}} \left(\frac{L_*}{L_{\odot}} \right)^{\frac{1}{4}} \left(\frac{r}{1 \text{ au}} \right)^{-\frac{1}{2}} \text{ K} \quad \text{--- (3)}$$

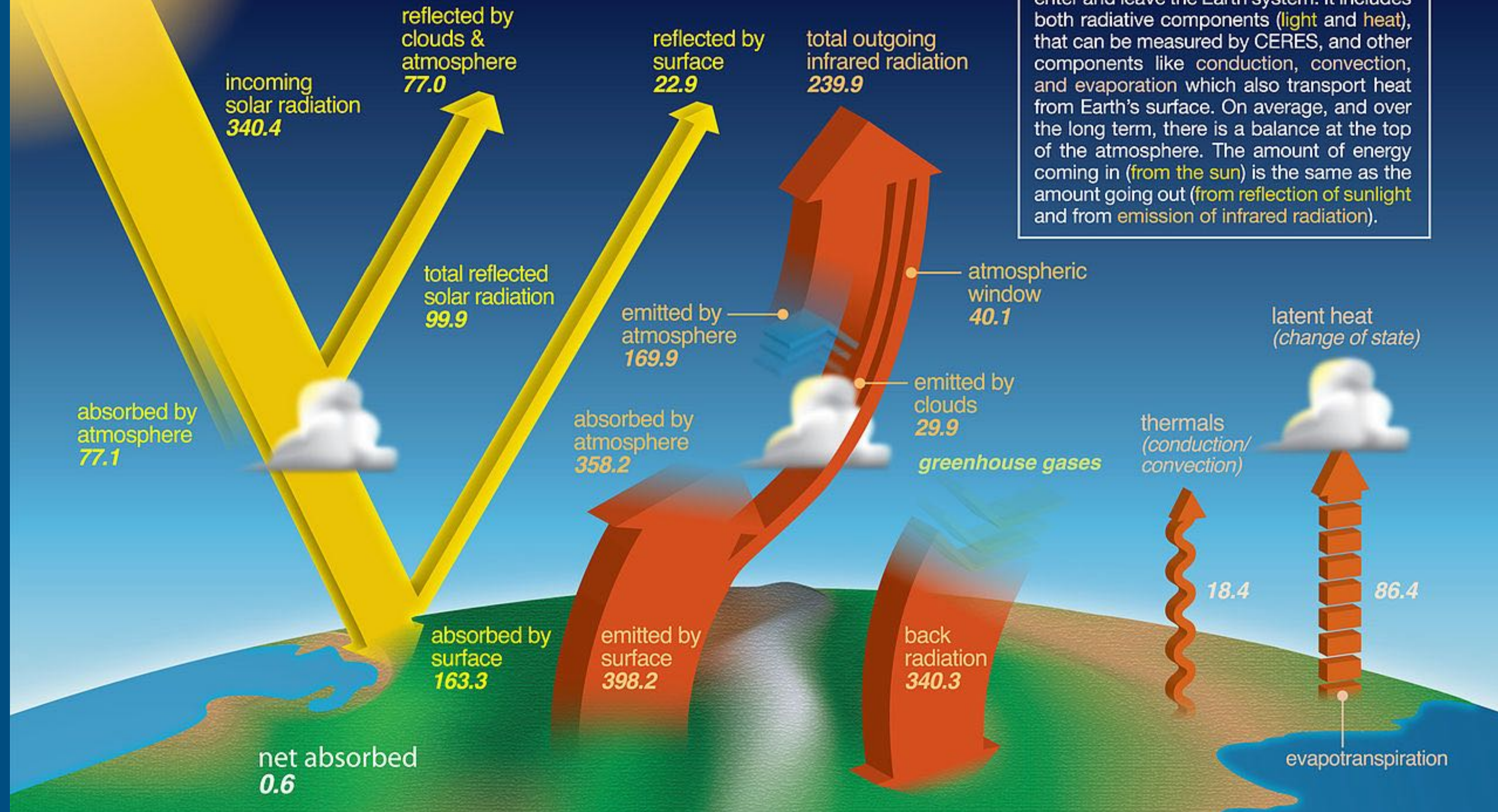
where $A_{\oplus} = 0.3$, $L_{\odot} = 3.83 \times 10^{26} \text{ W}$.

(Solar radiation flux at Earth's orbit \equiv the solar constant $S_{\odot} \equiv L_{\odot}/(4\pi r_{\oplus}^2) = 1360 \text{ W m}^{-2}$)



earth's energy *budget*

The Earth's energy budget describes the various kinds and amounts of energy that enter and leave the Earth system. It includes both radiative components (light and heat), that can be measured by CERES, and other components like conduction, convection, and evaporation which also transport heat from Earth's surface. On average, and over the long term, there is a balance at the top of the atmosphere. The amount of energy coming in (from the sun) is the same as the amount going out (from reflection of sunlight and from emission of infrared radiation).



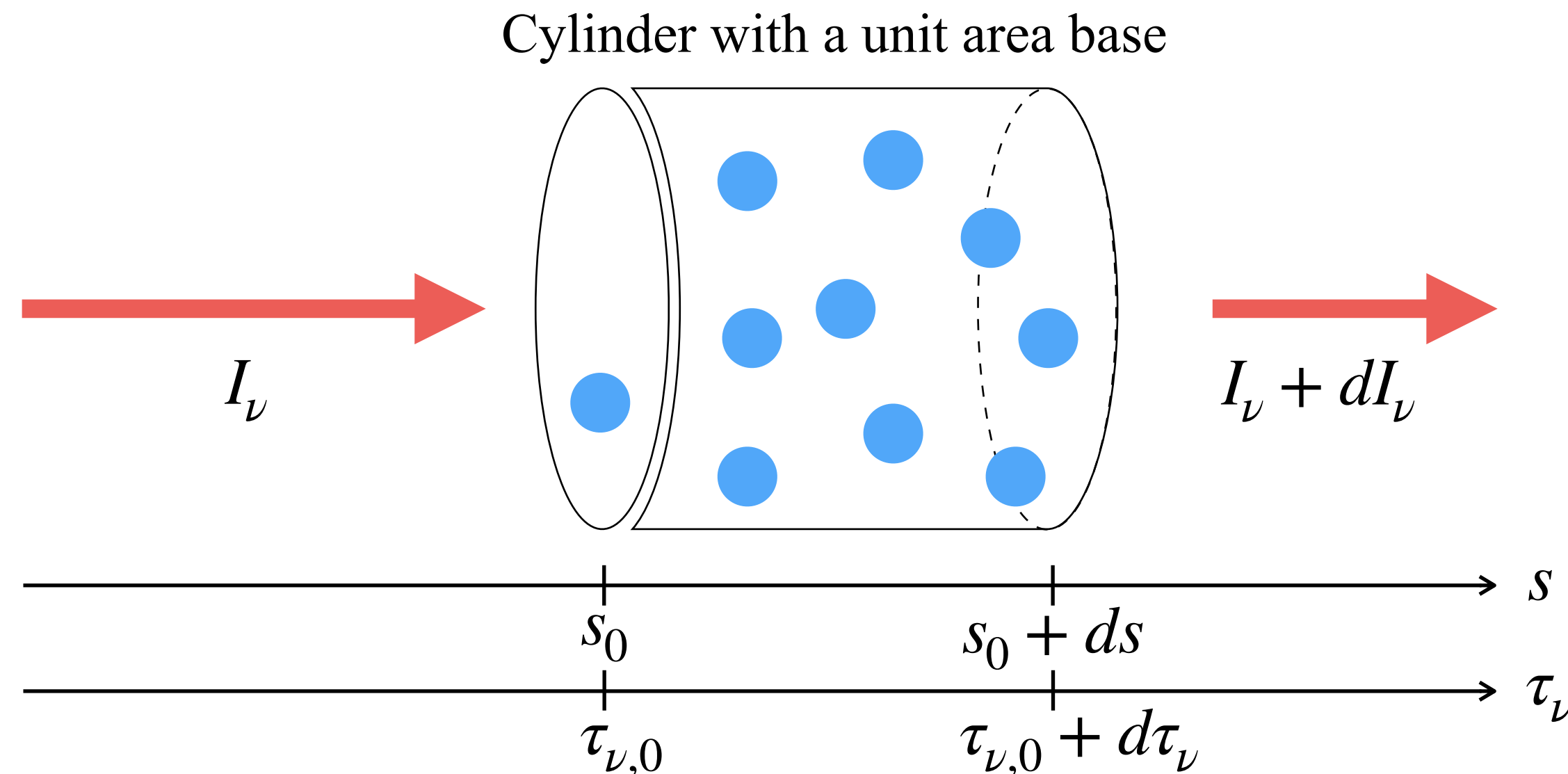
All values are fluxes in Wm^2
and are average values based on ten years of data

Loeb et al., J. Clim. 2009
Trenberth et al., BAMS, 2009

Optical depth

Intensity I_ν [$\text{W m}^{-2} \text{Hz}^{-1} \text{sr}^{-1}$]
 (Photon energy per unit time, area, frequency, and solid angle)

Absorber (e.g., atmospheric gas)
 Density: ρ_a [kg m^{-3}],
 Opacity (absorption cross-section per mass): κ [$\text{m}^2 \text{kg}^{-1}$]



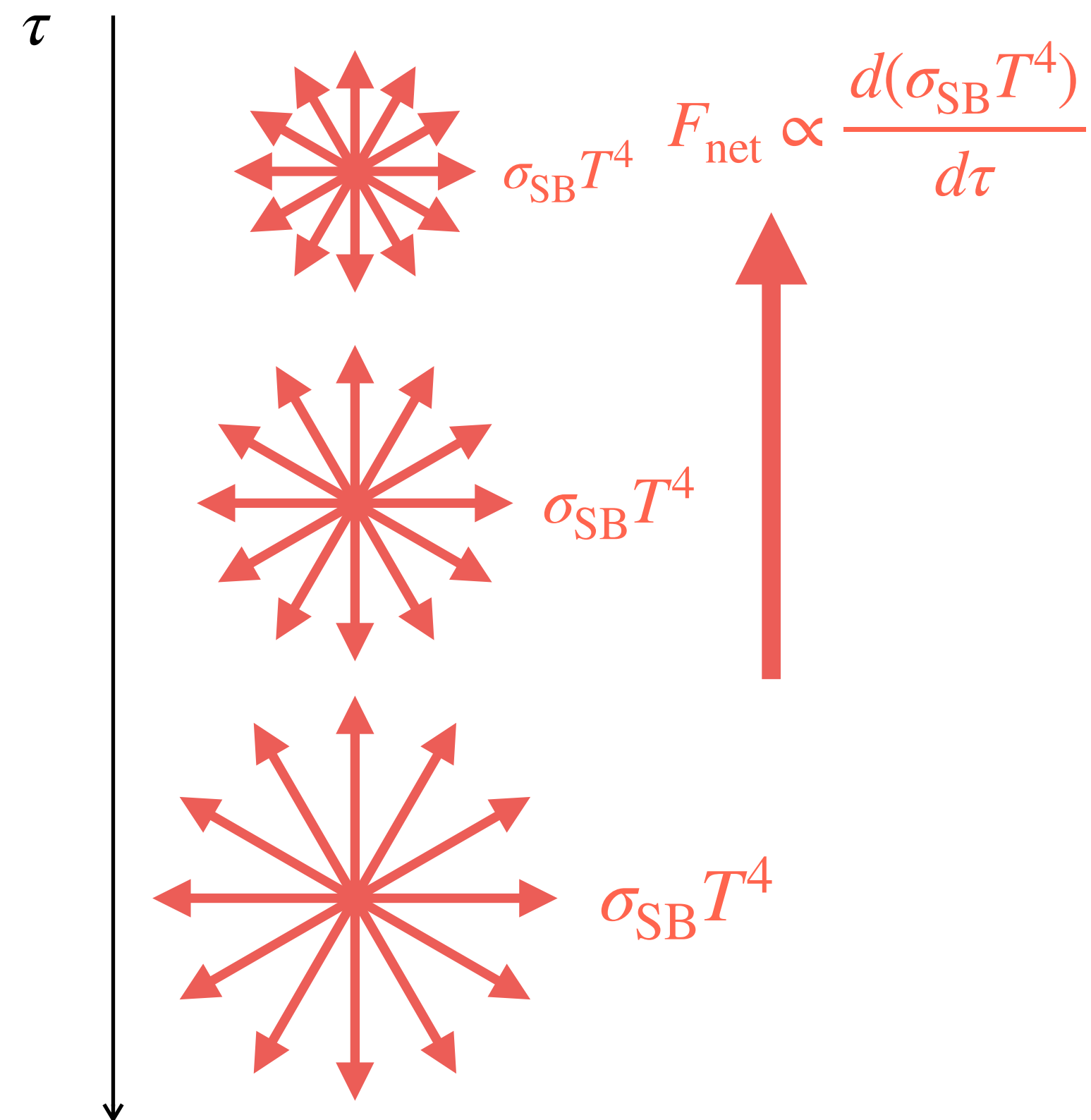
$$dI_\nu = -\kappa_\nu \rho_a ds \cdot I_\nu \quad (1).$$

Here we define *the optical depth*, $d\tau_\nu \equiv \kappa_\nu \rho_a ds$ — (2) and obtain,

$$\frac{dI_\nu}{d\tau_\nu} = -I_\nu \quad (3).$$

$\therefore I_\nu = I_\nu(s_0) \cdot \exp[-(\tau - \tau_0)]$ — (4) \rightarrow The radiation diminishes by $1/e$ every unit optical depth!

A simple radiative transfer model: diffusion approximation



Because each layer of the atmosphere emits the radiation $\sigma_{\text{SB}} T^4$, the radiation energy flux is given by,

$$F_{\text{rad}} = \frac{4}{3} \frac{d(\sigma_{\text{SB}} T^4)}{d\tau} \quad \text{--- (1)}$$

$$\text{As } d\tau = -\rho\kappa dz, \quad 4\sigma_{\text{SB}} T^3 \frac{dT}{dz} = -\frac{3}{4} F_{\text{rad}} \cdot \rho\kappa \quad \text{--- (2)}$$

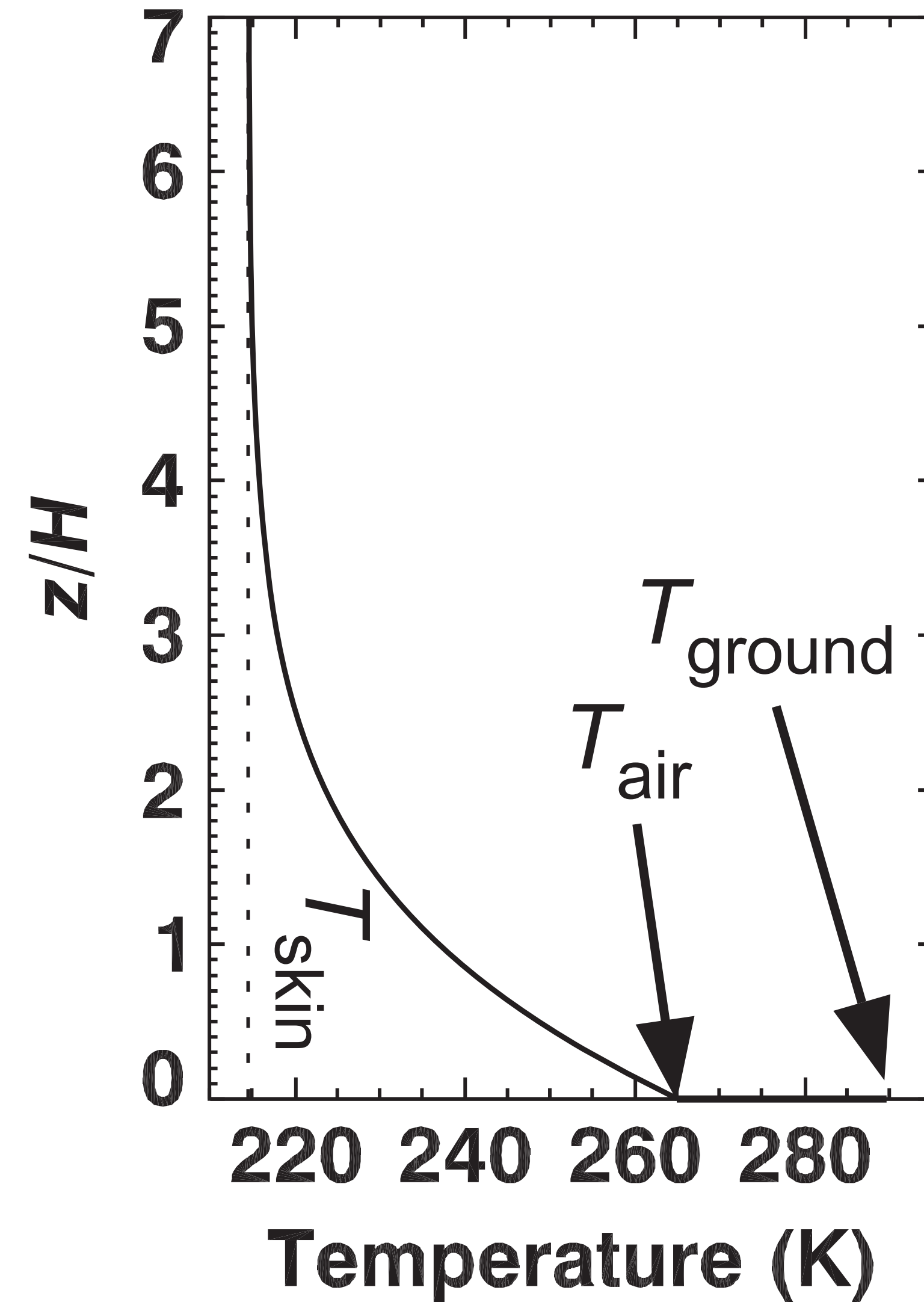
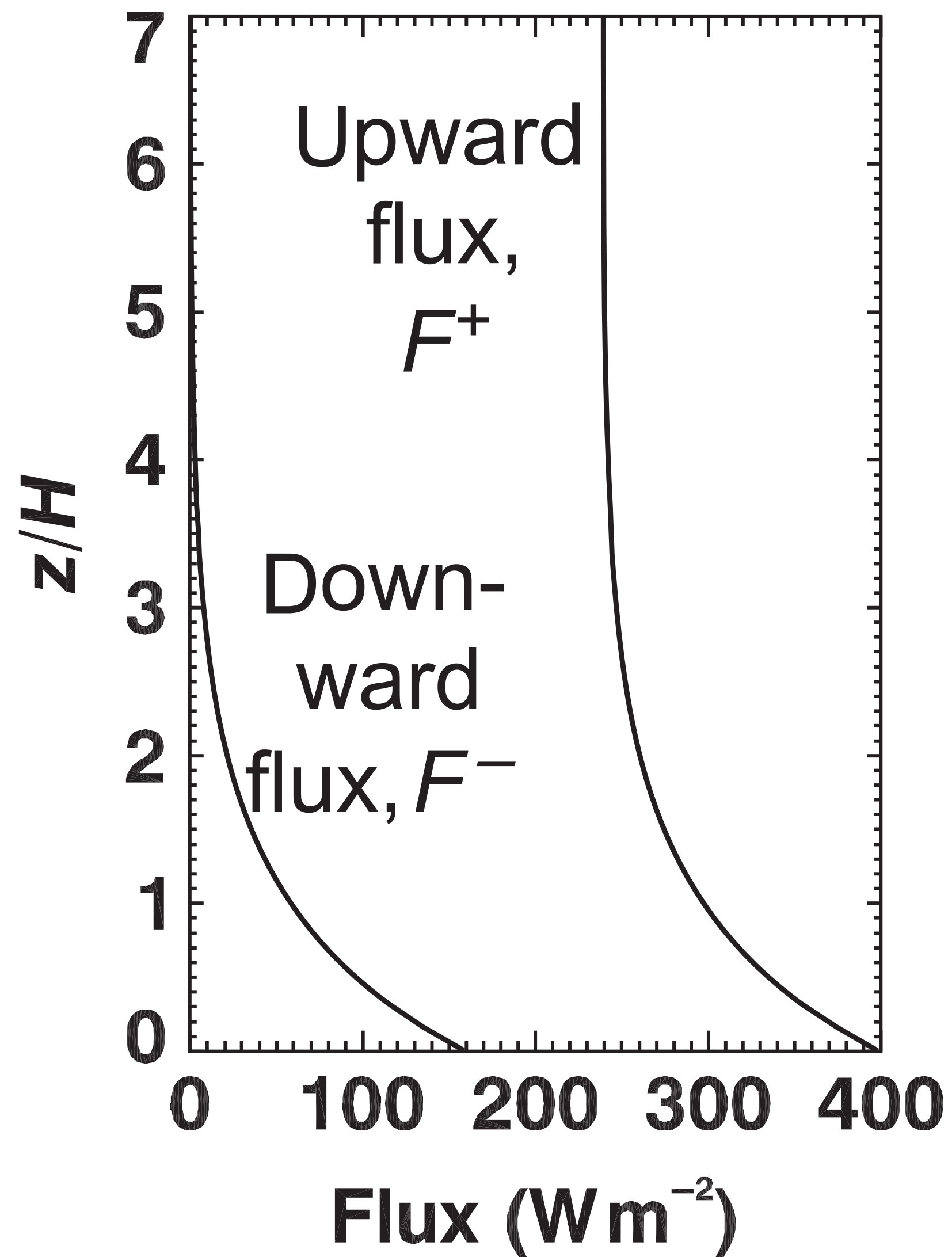
$$\therefore \left(\frac{dT}{dz} \right)_{\text{rad}} = -\frac{3\rho\kappa}{16\sigma_{\text{SB}} T^3} F_{\text{rad}} \quad \text{--- (3)}$$

This is called *the diffusion approximation of radiation*, which is valid for an optically-thick medium (i.e., $\tau \gg 1$).

Eq. 3 shows that energy is transported from high to low temperature layers.

For more precise derivation, see, for example, Rybicki & Lightman (1985), *Radiative Processes in Astrophysics*

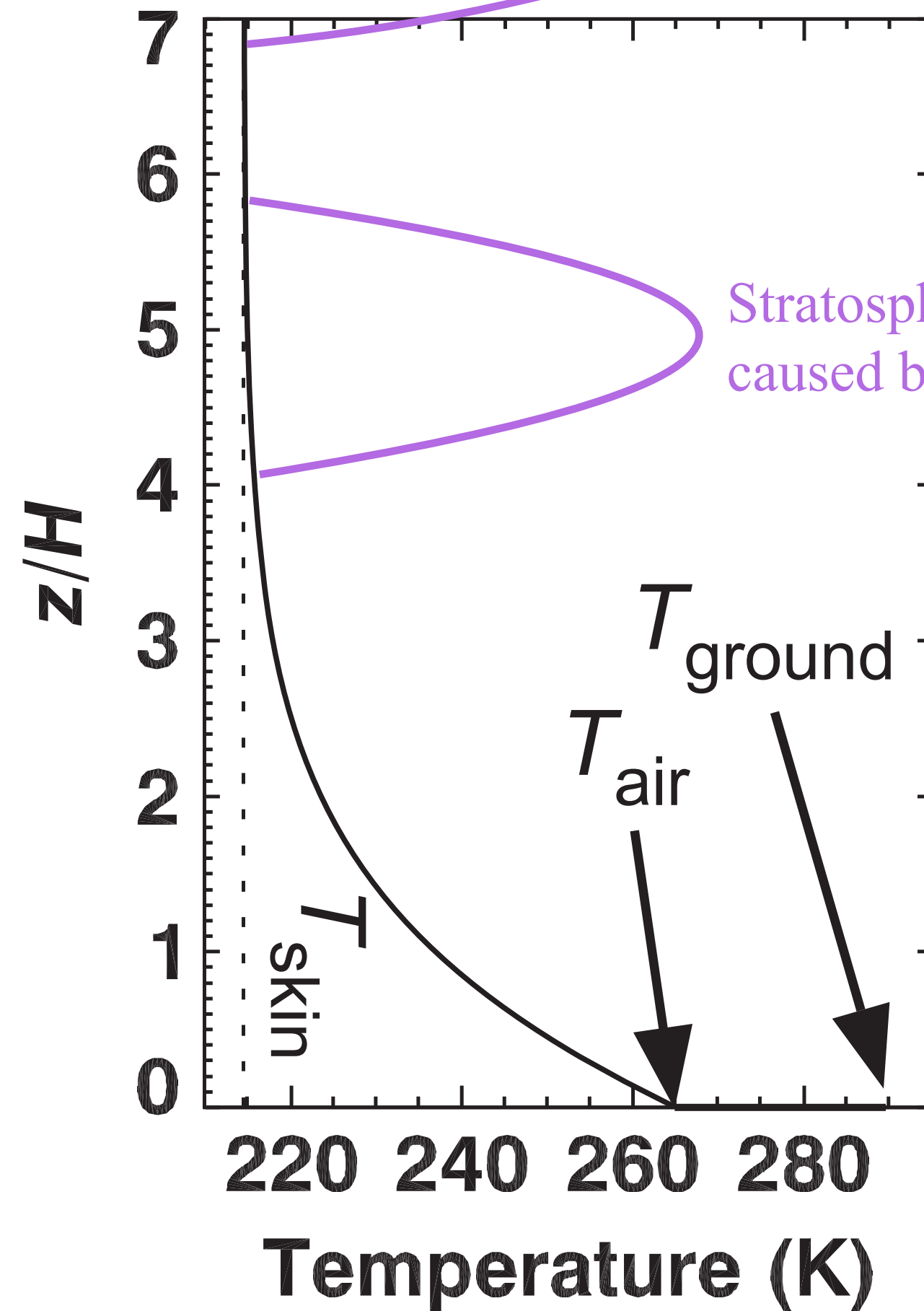
A radiative-equilibrium model for Earth's atmosphere



Catling & Kasting (2017)

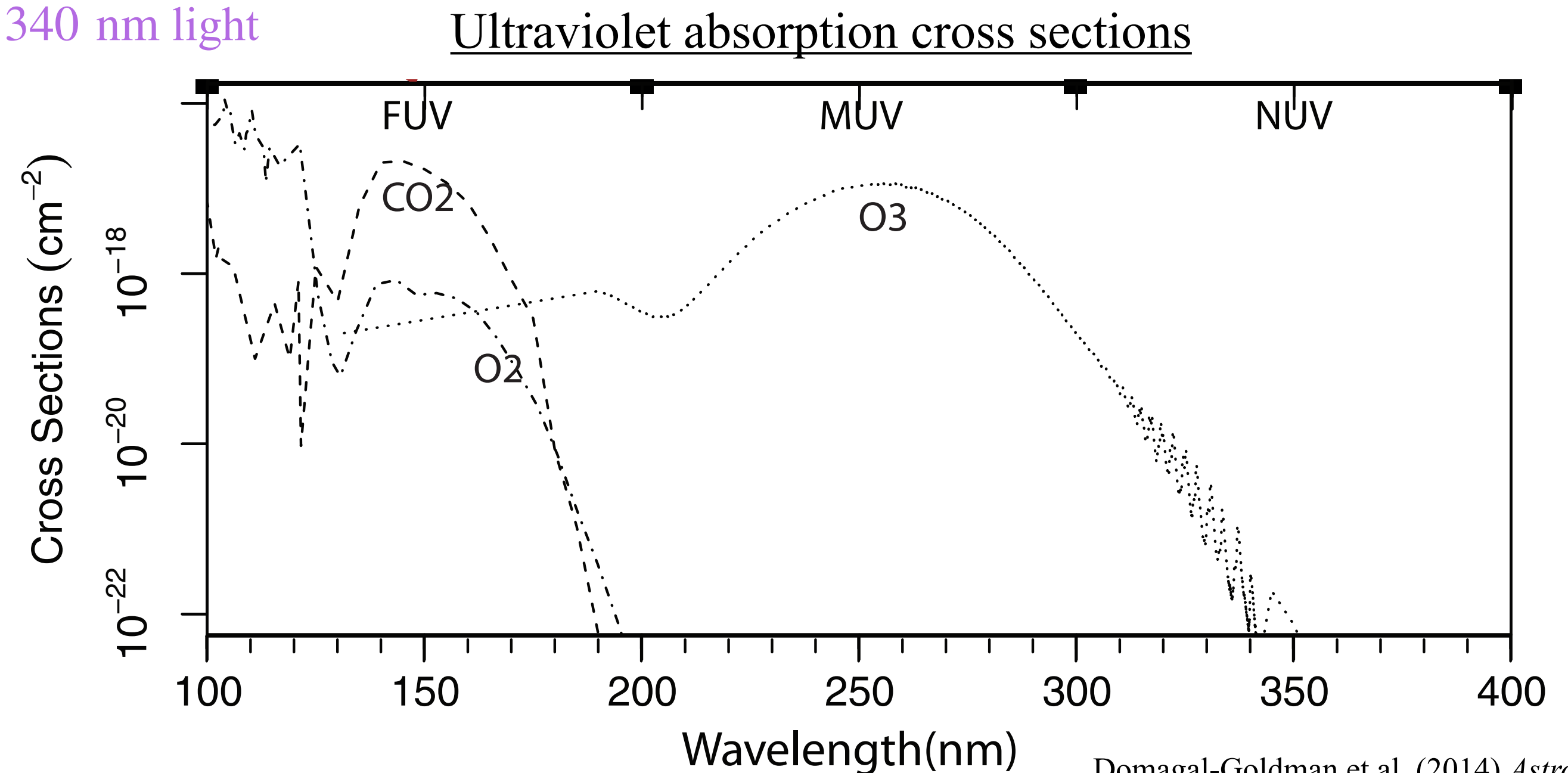
Heating by UV absorption

Thermospheric heating (common for all planets)
caused by absorbing $\lesssim 200$ nm light



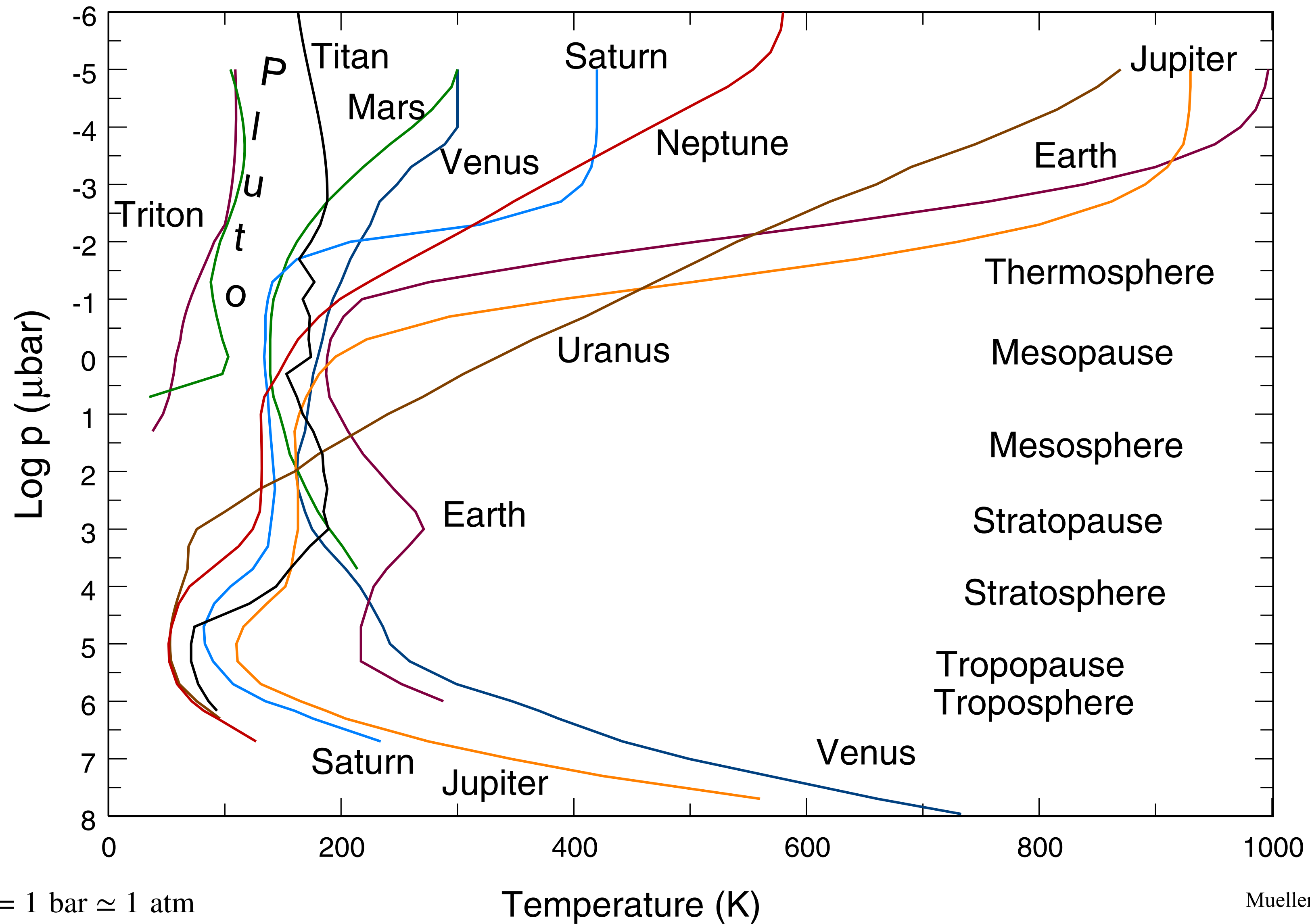
UV-absorption causes local heating in optically-thin (for infrared) layers

- Thermosphere: molecules such as CO_2 , O_2
- Stratosphere: O_3 (in the solar system, this is only for Earth!)
- Important for shielding DNA-damaging wavelengths ($\lesssim 300$ nm)



Domagal-Goldman et al. (2014) *Astrophys. J.*

Atmospheric temperature profiles of the solar system bodies

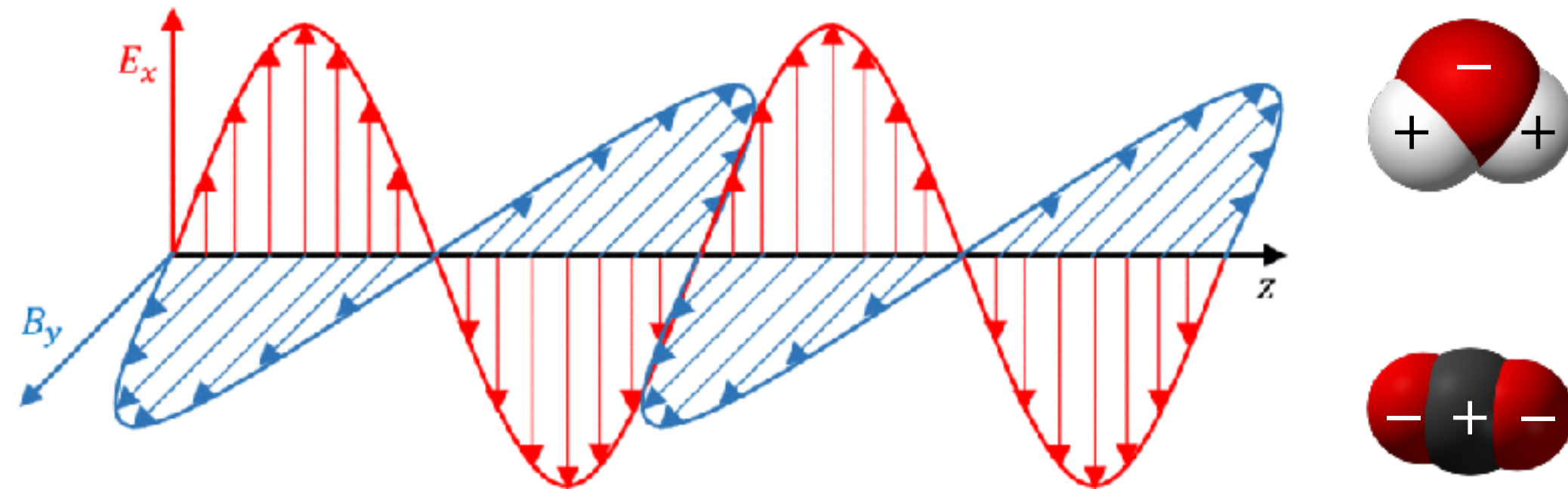


Note: $10^5 \text{ Pa} = 1 \text{ bar} \simeq 1 \text{ atm}$

Mueller-Wodarg et al. (2008)

How do atmospheric gases absorb light?

- Absorption and emission of photons = Interactions of molecules/atoms with electromagnetic wave
 - Greenhouse gases : H₂O (structural electric dipole moment), CO₂ (stretch- and bending-induced)
 - Non-green house gases : N₂ (no dipole moment), O₂ (has a magnetic dipole but in radio wavelengths)
- The energy absorbed = Transition energy between different energy levels (absorption line)

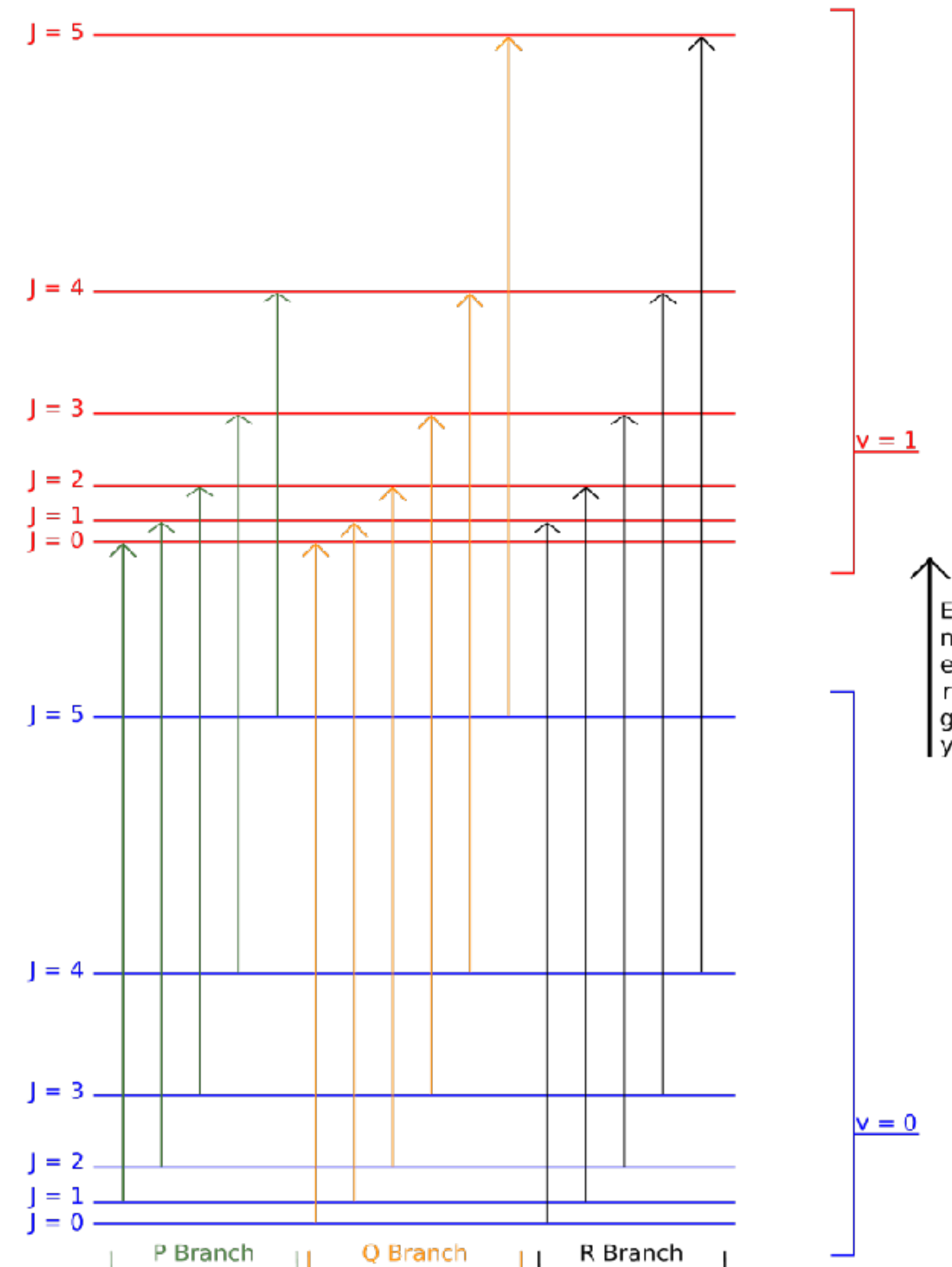
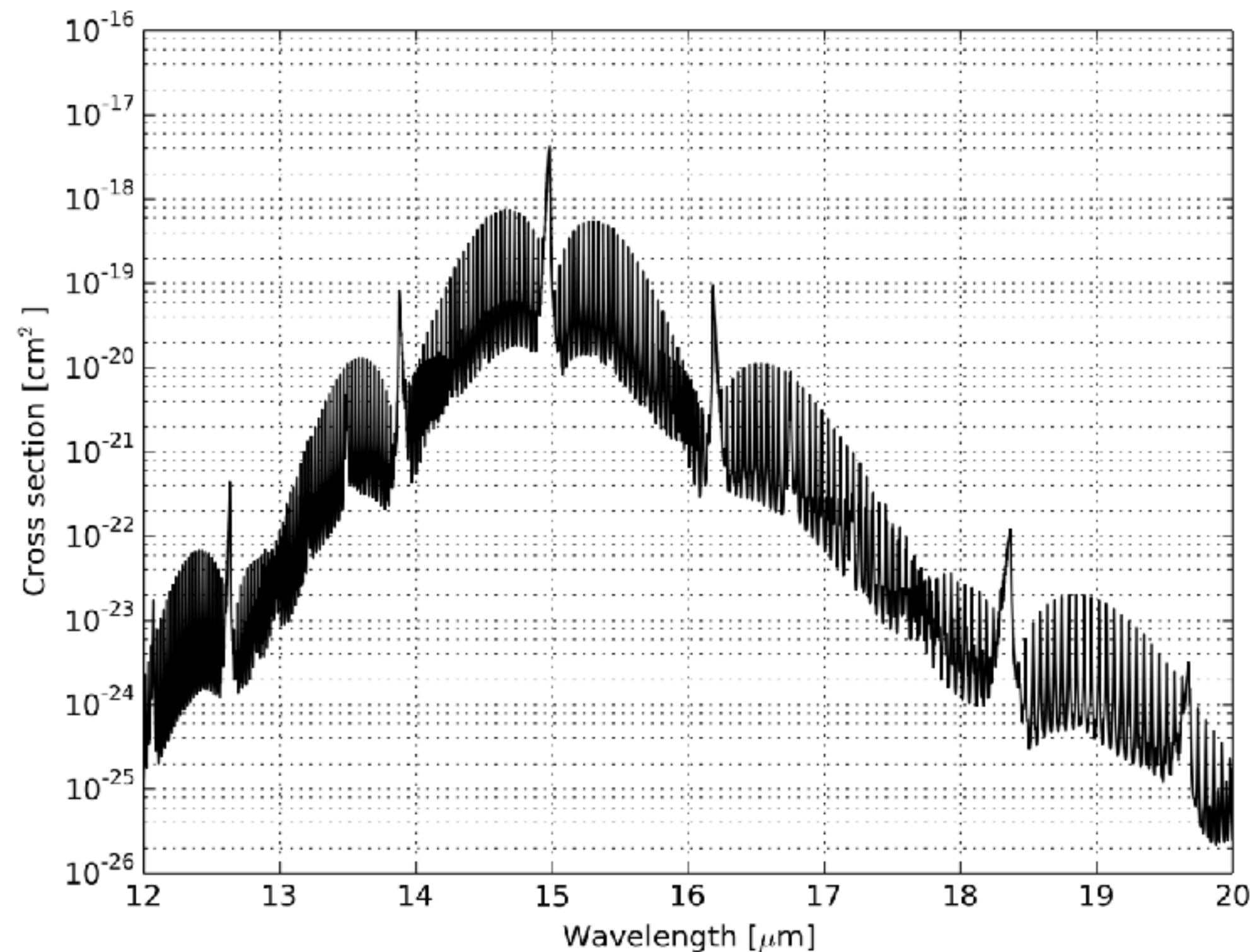


Transition	Wavelength
Electron	< 1 μm
Vibration	1-20 μm
Rotation	> 20 μm

Absorption band

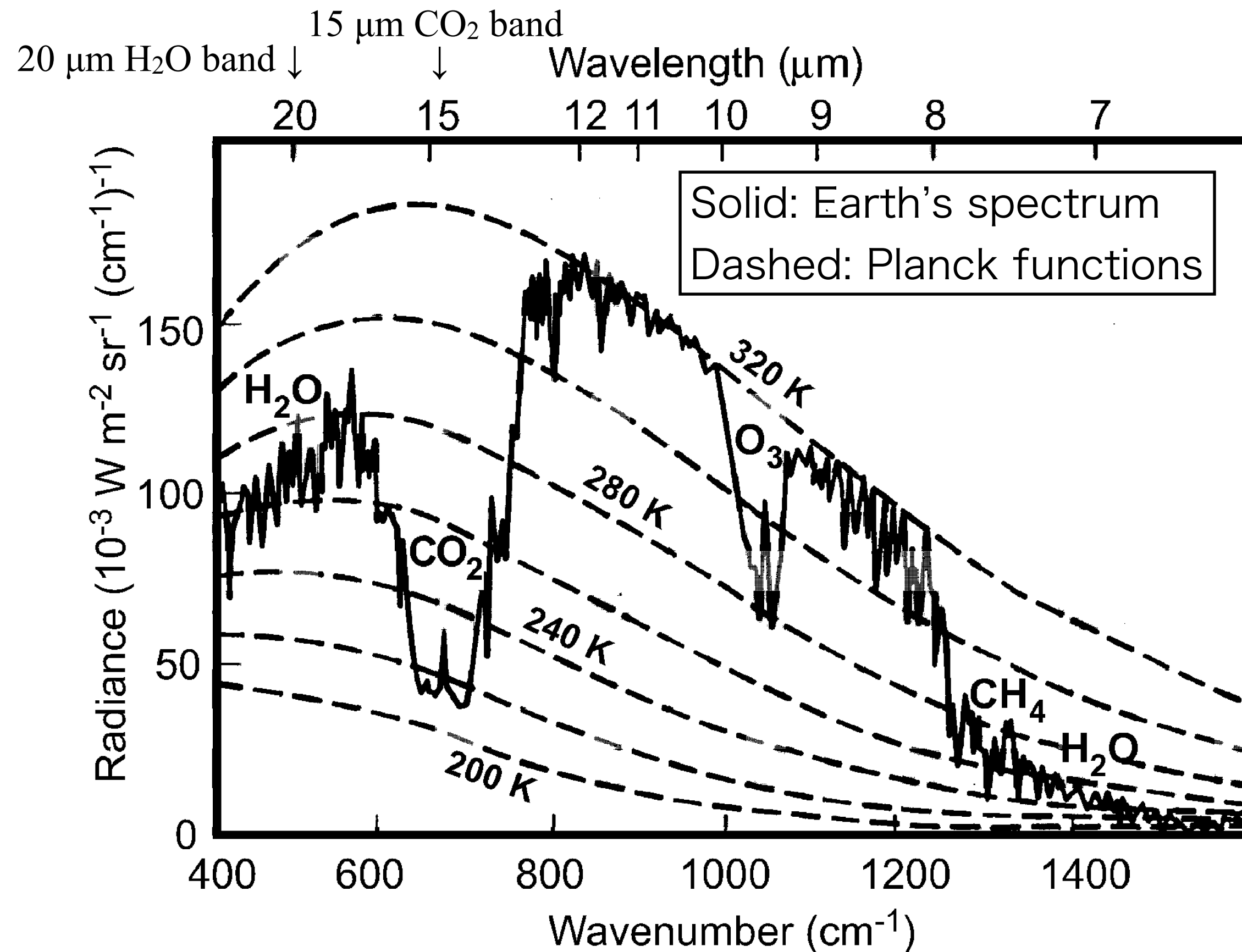
In infrared wavelengths, vibrational transition (wide spacing) combined with rotational transition (narrow) form absorption bands (*vib-rotational transition*)

15- μm absorption band of CO₂ (from HITRAN database)



Atmospheric window

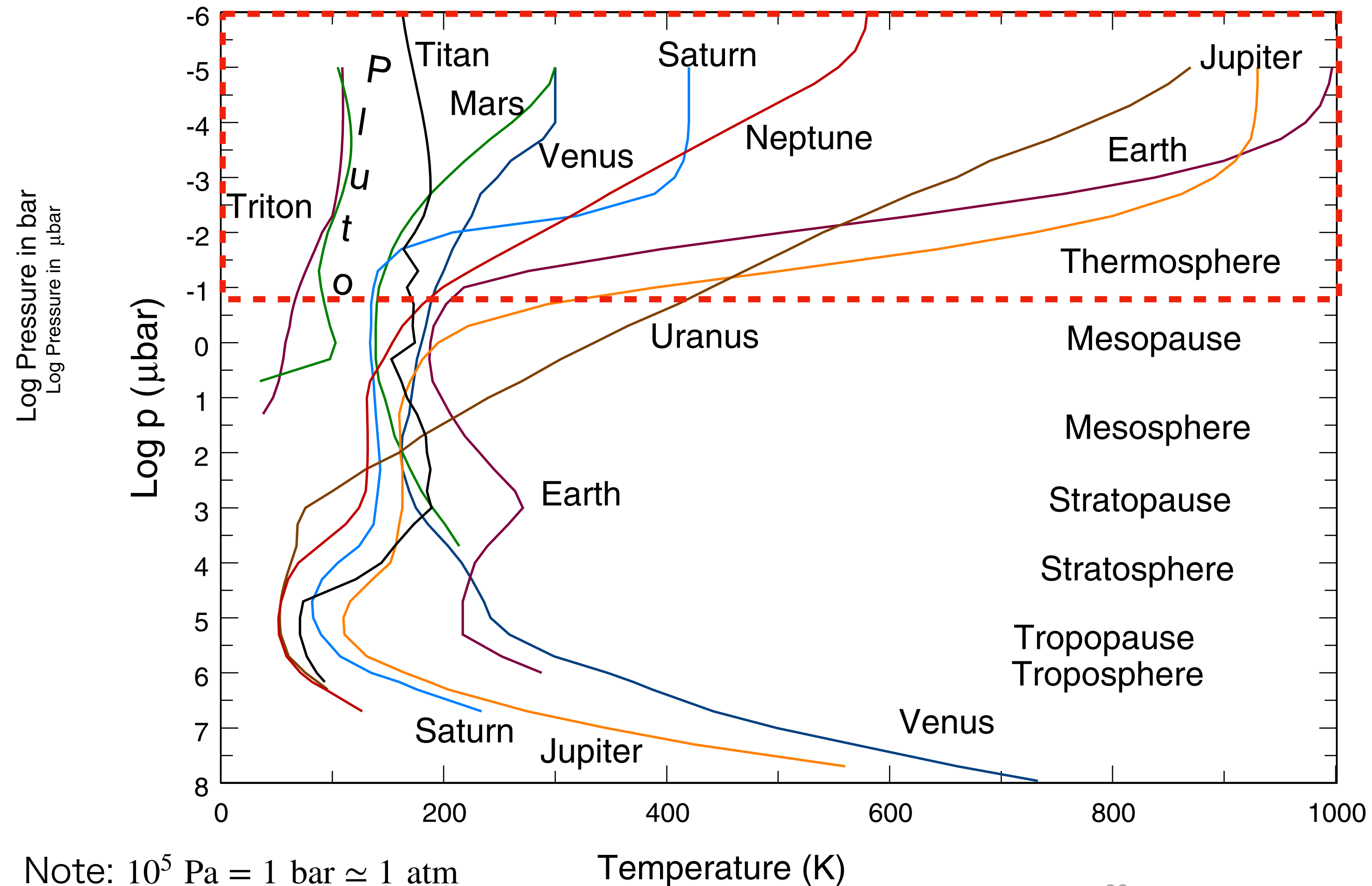
Earth's emission spectrum



- Infrared absorption/emission is wavelength dependent
- Less emission in wavelengths where absorption bands exist
 - Emission from the surface is absorbed in the atmosphere
 - Emission from $\tau_\nu = 1$ layer reaches the space
- **Atmospheric window:** optically-thin wavelengths
 - If multiple greenhouse gases have absorption bands in window wavelengths of each other, their effects are maximized
 - Note that the atmospheric windows are important for ground-based observations

Kirchhoff's law of thermal radiation

**For a material body in thermodynamic equilibrium,
absorptivity = emissivity**



**Low temperature thermospheres
of Venus and Mars due to
 CO_2 cooling**

Note: $10^5 \text{ Pa} = 1 \text{ bar} \simeq 1 \text{ atm}$

Temperature (K)

Tropopause visualized by a cumulonimbus cloud



The derivation (for physics students)

Let us derive the adiabatic lapse rate, defined by,

$$\Gamma_a \equiv - \left(\frac{dT}{dz} \right)_a \quad \text{--- (1)}.$$

Here we consider a unit-mass air parcel moving upward/downward adiabatically (no energy exchange).

From the first law of thermodynamics,

$$0 = dq = c_v dT + pd(1/\rho) \quad \text{--- (2)}.$$

From the ideal gas law, $p = \rho \bar{R}T$ --- (3), we obtain,

$$d(1/\rho) = (\bar{R}/p)dT - (\bar{R}T/p^2)dp \quad \text{--- (4)}.$$

Substituting Eq. 4 into Eq. 2 gives,

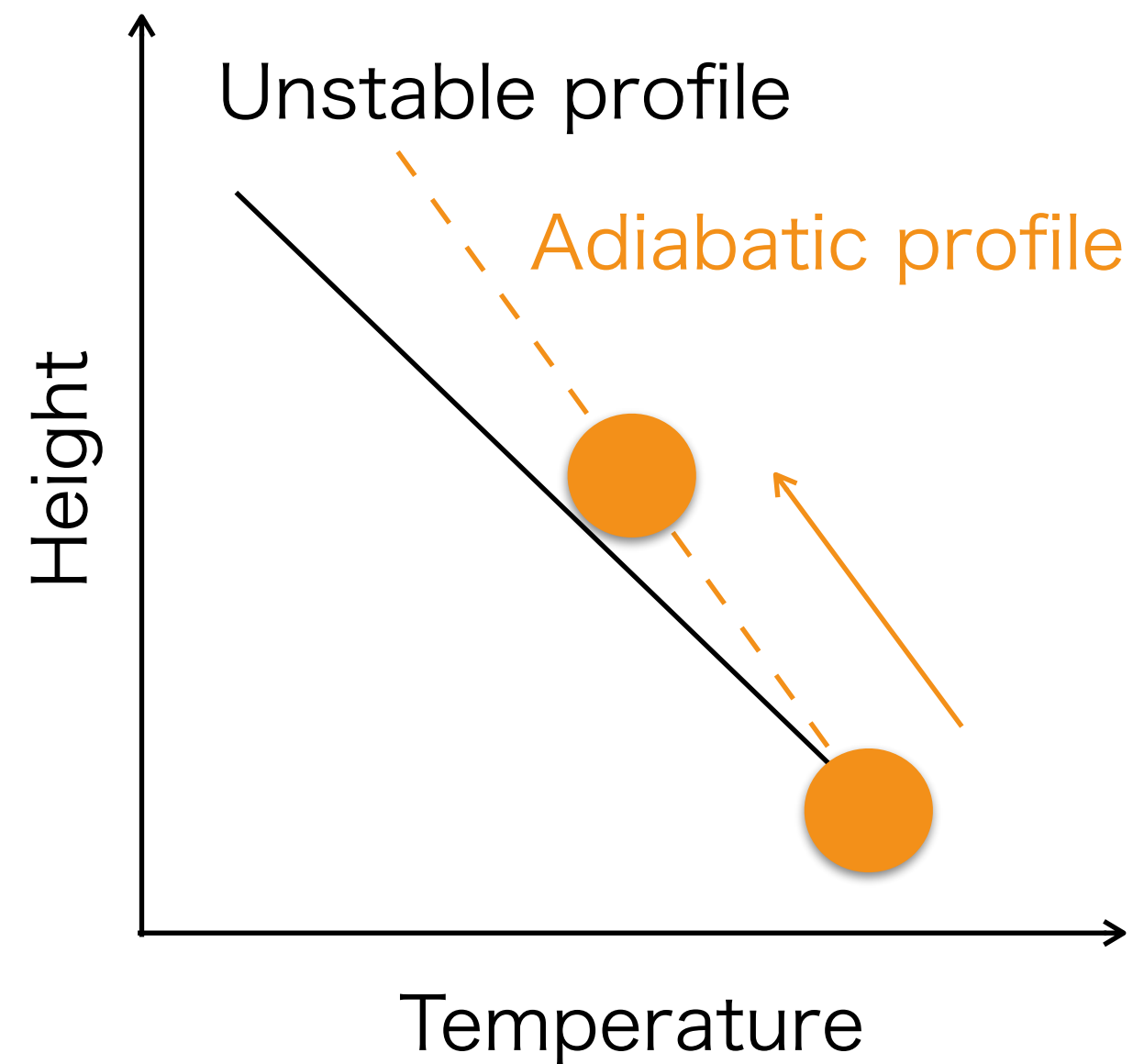
$$0 = (c_v + \bar{R})dT - (\bar{R}T/\rho)dp \quad \text{--- (5)}.$$

Next we substitute the hydrostatic equation, $dp/dz = -\rho g$ --- (6), into Eq. 5 and obtain,

$$0 = c_p dT + g dz \quad \text{--- (7)}.$$

$$\therefore \Gamma_a \equiv - \left(\frac{dT}{dz} \right)_a = \frac{g}{c_p} \quad \text{--- (8)}.$$

Adiabatic temperature lapse rate



The air parcel is hotter than surrounding

→ lower density

→ Positive buoyancy

- When the radiative-equilibrium temperature profile is unstable against convection, the atmosphere start to convect

- Because the efficiency of energy transport by convection is efficient, the temperature profile in the convective layer is given by *the adiabatic temperature lapse rate*.

Given $C_p = 1.0 \times 10^3 \text{ J K}^{-1} \text{ kg}^{-1}$, $g = 9.8 \text{ m s}^{-2}$,

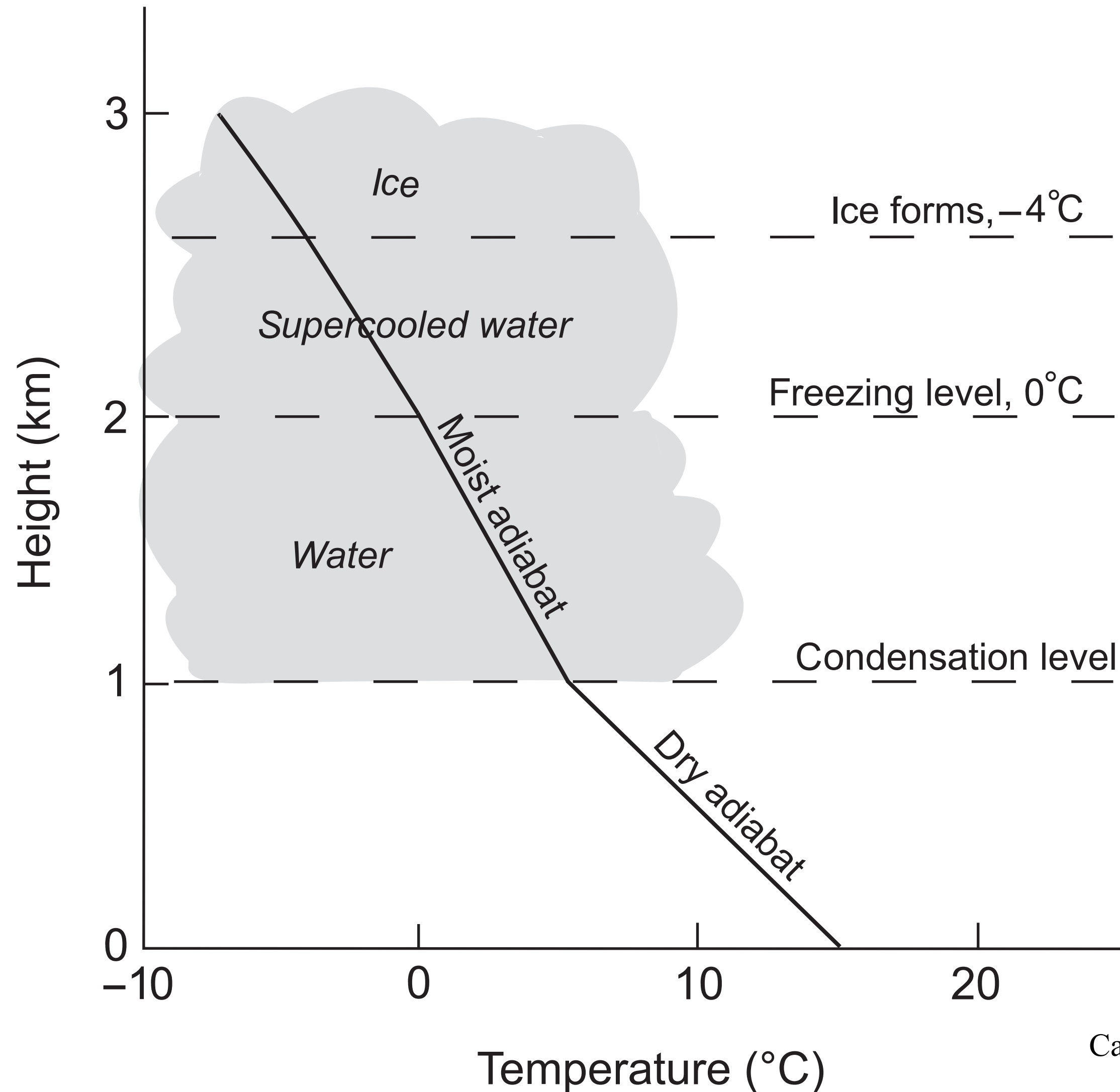
$$\Gamma_a \equiv - \left(\frac{dT}{dz} \right)_{\text{adiabatic}} = \frac{g}{C_p} \simeq 10 \text{ K/km} \quad \text{--- (1)}$$

- In other words, the criterion for convection is given by,

$$- \left(\frac{dT}{dz} \right) > - \left(\frac{dT}{dz} \right)_{\text{adiabatic}} \quad \text{--- (2)}$$

Eq. 2 is an approximated form of the Rayleigh criterion.

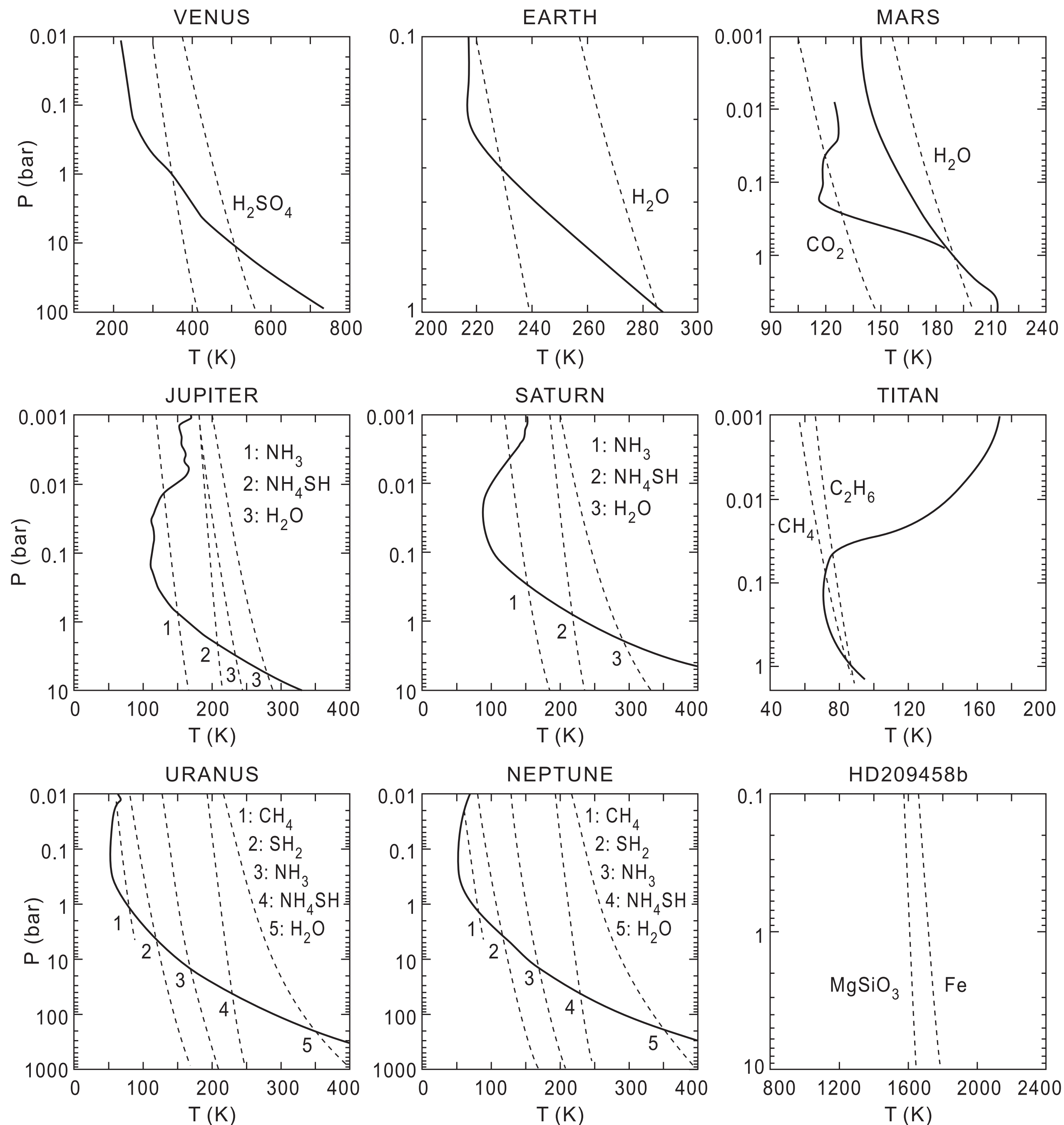
Moist convection



Catling & Kasting (2017)

- While the adiabatic lapse rate is estimated as
$$\Gamma_a \equiv - \left(\frac{dT}{dz} \right)_{\text{adiabatic}} = \frac{g}{C_p} \simeq 10 \text{ K/km} \quad \text{--- (1),}$$
the actual lapse rate in Earth's troposphere is
$$\simeq 6 \text{ K/km} \quad \text{--- (2)}$$
- The difference is caused by *moist convection*, (the lapse rate is called *moist adiabat*) where condensed water releases the latent heat and warm up the updraft gas

Clouds



Solid: atmospheric profiles
 Dashed: Saturated vapor pressure curves
 (Figure from Catling & Kasting, 2017)

- In general, tropospheric lapse rate is steeper than saturated vapor pressure curves
 → Cloud formation in a updraft
- Condensable gases dependent on the temperature and composition

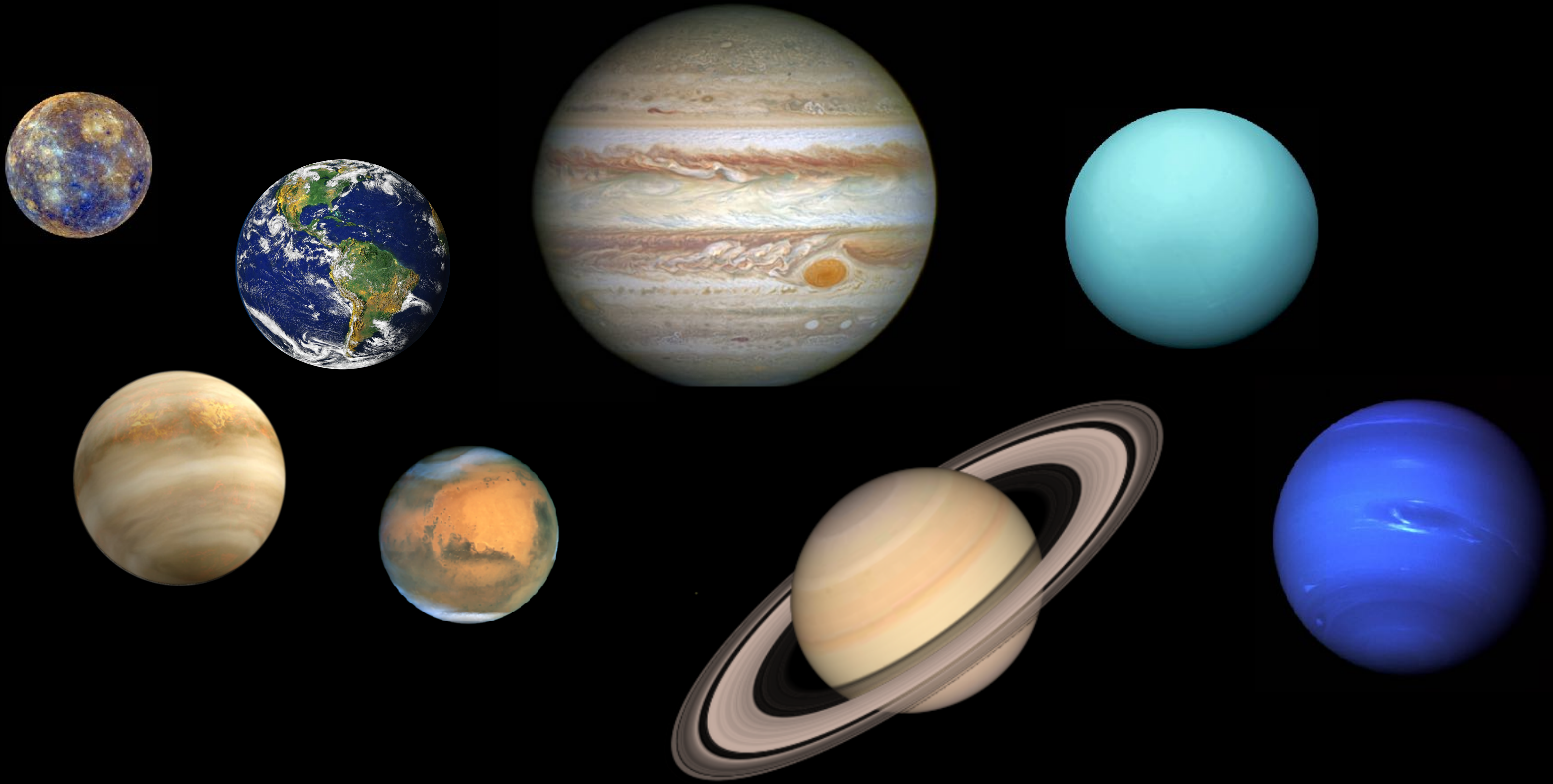
e.g.,

Venus: H₂SO₄

Earth: H₂O

Mars: H₂O, CO₂

Note: $10^5 \text{ Pa} = 1 \text{ bar} \simeq 1 \text{ atm}$



Equilibrium chemistry

Here we consider a hypothetical reaction $bB + cC \rightleftharpoons gG + hH$ — (1),

Whose equilibrium state is given by the equilibrium constant K_{eq} , given as,

$$K_{\text{eq}} = \frac{a_G^g \cdot a_H^h}{a_B^b \cdot a_C^c} \quad \text{— (2)}$$

where a is the activity (effective concentration considering non-ideal effects).

For ideal gases, a is given by the partial pressure p .

The equilibrium constant K_{eq} is related with Gibbs free change ΔG .

$$K_{\text{eq}} = \exp\left(-\frac{\Delta G}{RT}\right) \quad \text{— (3).}^\dagger$$

Once ΔG is given (for typical reactions one can find tables), the equilibrium abundances can be computed.

[†]One can obtain this relation by substituting the definition of Gibbs free energy into Eq. 2

Is Earth's atmosphere in chemical equilibrium?

Earth's atmosphere contains 0.21 bar O₂ (oxidizing) と 1.8 ppmv CH₄ (reducing)...

Here we consider their reaction, $\text{CH}_4 + 2\text{O}_2 \rightleftharpoons \text{CO}_2 + 2\text{H}_2\text{O}$ — (4)

Under the ambient temperature (25°C), $\Delta G = -817 \text{ kJ mol}^{-1}$ — (5)

Substituting (5) into (2) and (3), we obtain,

$$p_{\text{CH}_4} = \frac{p_{\text{CO}_2} \cdot a_{\text{H}_2\text{O}}}{p_{\text{O}_2} \cdot K_{\text{eq}}} = \frac{380 \times 10^{-6} \times 1}{(0.21^2 \times 10^{143.29})} = 10^{-145} \text{ bar !} \text{ — (6)}$$

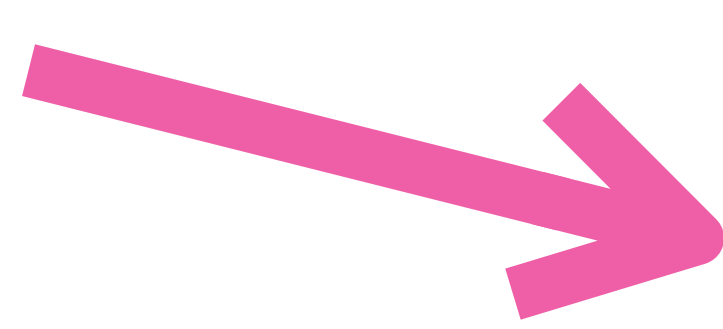
∴ Earth's atmosphere is in a non-equilibrium state.

Non-equilibrium chemistry driven by photodissociation

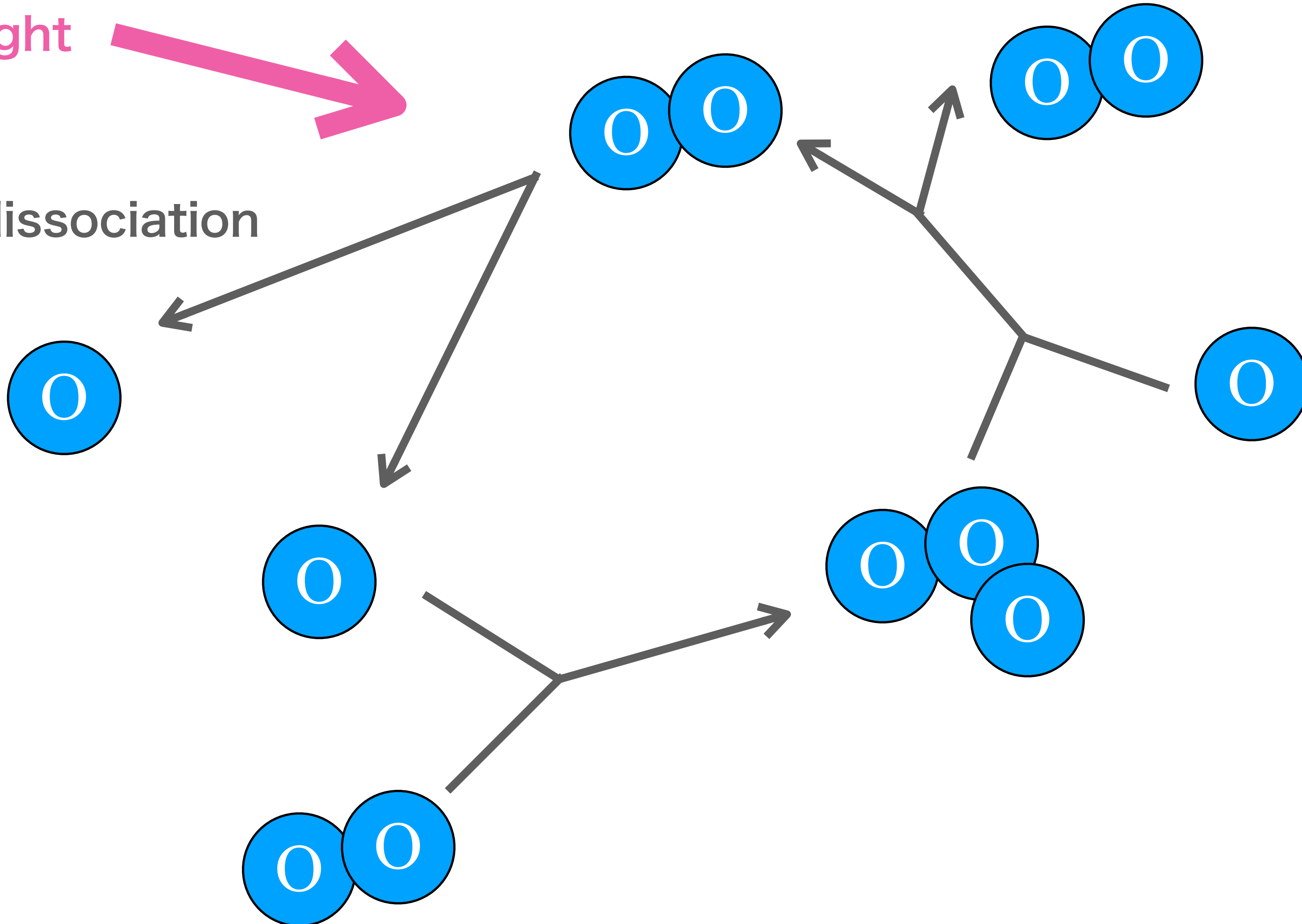
- **Radicals** produced by photodissociation drive non-equilibrium chemistry
 - Species with unpaired electrons in the outermost shell: such as **OH, Cl, O**
- Production of **OH**
 - Earth: $\text{O}_3 + h\nu (\lambda < 310 \text{ nm}) \rightarrow \text{O}_2 + \text{O}(^1D) \text{ — (1)}$, $\text{H}_2\text{O} + \text{O}(^1D) \rightarrow \text{OH} + \text{OH} \text{ — (2)}$
 - Mars: $\text{H}_2\text{O} + h\nu (\lambda < 240 \text{ nm}) \rightarrow \text{OH} + \text{O} \text{ — (3)}$
- Free energy of **radicals** obtained from photons propagates through reactions
 - e.g., $\text{CH}_4 + \text{OH} \rightarrow \text{CH}_3 + \text{H}_2\text{O} \text{ — (4)}$
- Eventually thermalized either by disproportionation reaction or recombination by three-body reaction.
 - e.g., $\text{OH} + \text{HO}_2 \rightarrow \text{H}_2\text{O} + \text{O}_2 \text{ — (5)}$, $\text{NO}_2 + \text{OH} + \text{M} \rightarrow \text{HNO}_3 + \text{M} \text{ — (6)}$

Photochemical production of ozone

Ultraviolet light



Photodissociation



OH — “Detergent of the atmosphere”

Trace gases are oxidized and removed from Earth’s atmosphere by OH

- $\text{CO} \rightarrow \text{CO}_2$ (~3 months)
- $\text{C}_x\text{H}_y \rightarrow \text{CO}_2$ (e.g., CH_4 ~10 years)
- $\text{N}_x\text{O}_y \rightarrow$ nitric acid
- SO_2 , H_2S , COS , dimethyl sulfide (CH_3SCH_3) \rightarrow sulfuric acid aerosols

On a planet without water, oxidation by OH (+ dissolution in rainfall) would not work, resulting in a completely different atmospheric composition!

e.g,) SO_2 gas in exoplanet atmospheres \rightarrow absence of liquid water? (Luftus et al. 2019, *Astrophys. J.*)

Summary

- The solar system planets show a diversity in their atmospheres
- Pressure and density drop exponentially with height
 - Atmospheric gas can escape to space from the exosphere
- The greenhouse effect is important to determine the surface temperature
 - Due to radiative transfer properties, the lower atmosphere is hotter
- UV-absorption causes additional heating in the thermosphere and stratosphere
- The troposphere is convective
 - The temperature lapse rate is determined by (moist) adiabat
 - Cloud formation depending on the temperature and composition
- Atmospheres are in chemical dis-equilibrium
 - Biological processes
 - Photochemistry — ozone production, OH reactions

Report assignment

Summarize your answers into a short report and submit it by the beginning of the next lecture (either directly, to my post-box, or by e-mail to hiro.kurokawa@elsi.jp).

1. Given the mean surface temperature, $T = 288$ K, and the mean molecular mass, $\bar{m} = 29.0 \times 10^{-3} / 6.02 \times 10^{23}$ kg, estimate the scale height of Earth's atmosphere. Answer with two significant digits.

$$H = \frac{k_B T}{\bar{m} g} \simeq \boxed{} \text{ km.}$$

2. Estimate the pressure at the top of Mt. Everest (8,849 m). You can use the pressure at the sea level = 1.013×10^5 Pa, and the scale height from Q1. Answer with two significant digits.