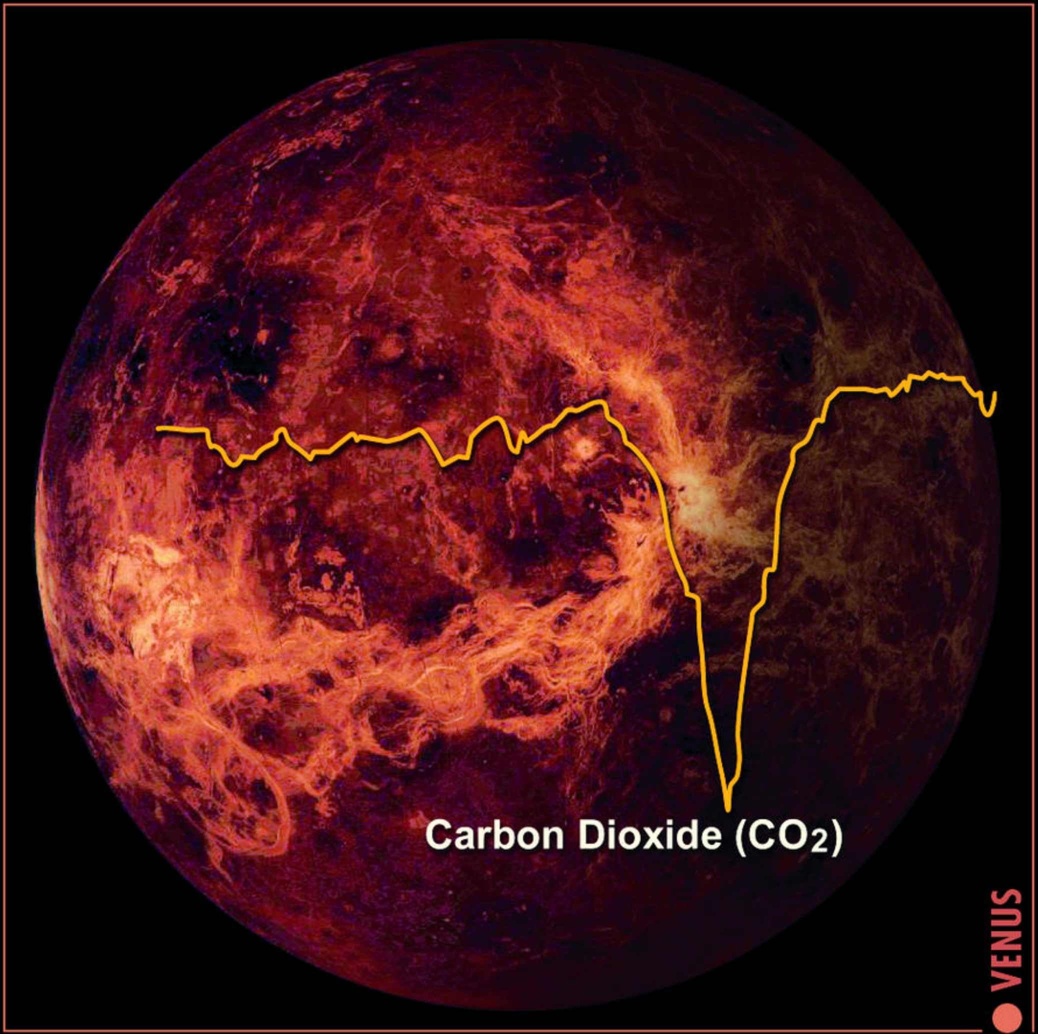
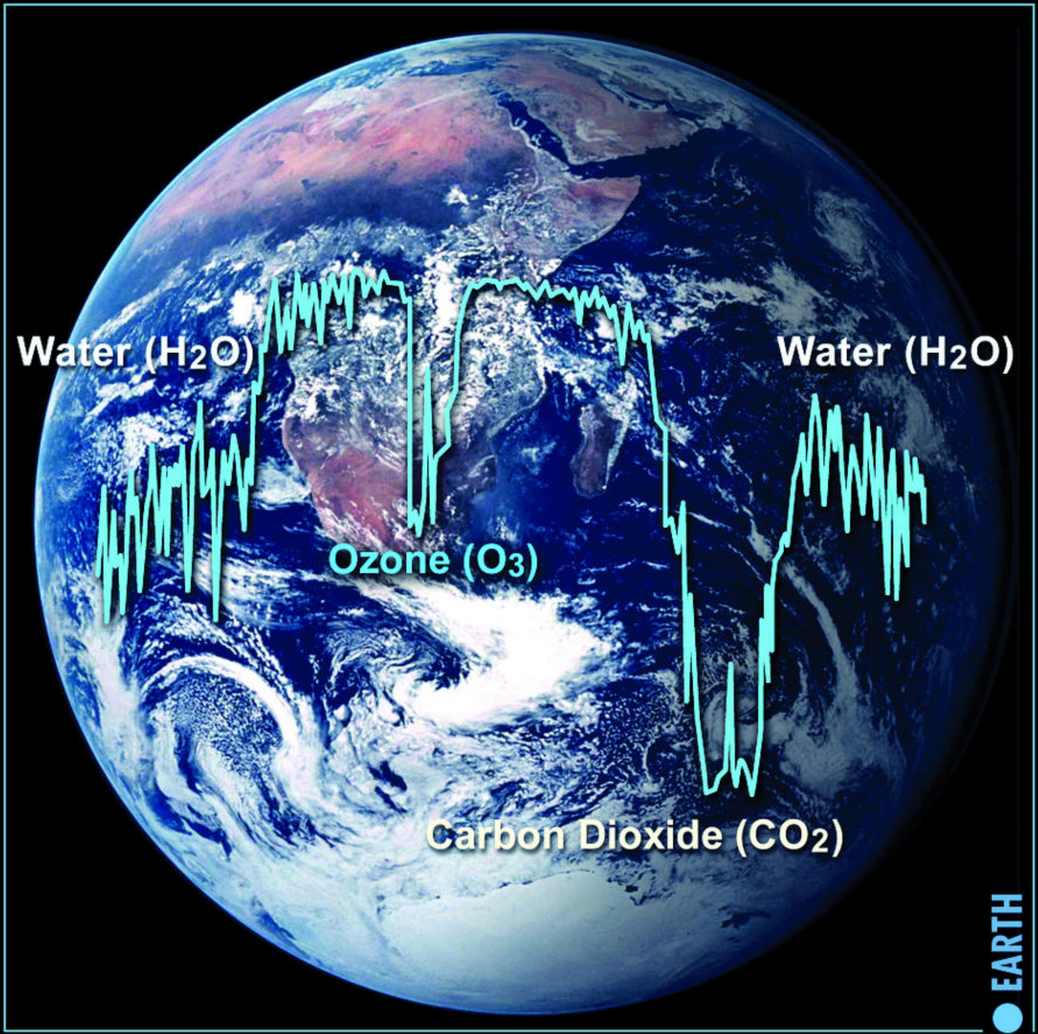
Planetary Habitability





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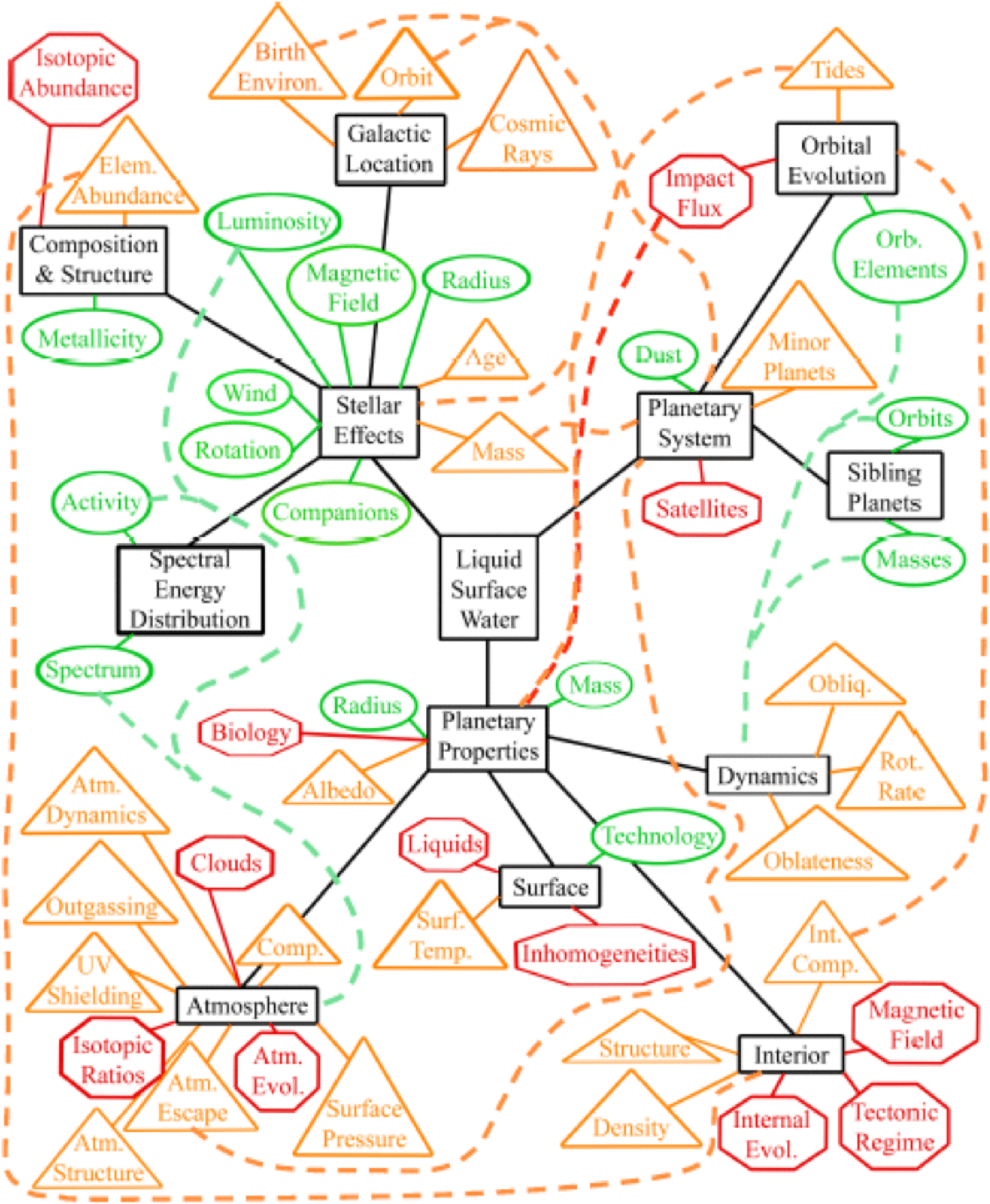
Topics

- Lecture 1 – Introduction
- Lecture 2 – Habitability Factors
- Lecture 3 - Stars
- Lecture 4 – Planetary Atmospheres
- Lecture 5 – Planetary Interiors
- Lecture 6 – Planetary Energy Balance
- Lecture 7 – Habitable Zone I
- Lecture 8 – Habitable Zone II
- Lecture 9 – Earth as a Living Planet
- Lecture 10 – Mars
- Lecture 11 – Icy Moons
- Lecture 12 – Venus
- Lecture 13 – Mercury & the Moon
- Lecture 14 – The Role of Giant Planets
- Lecture 15 – Stellar Influences
- Lecture 16 – Magnetic Fields
- Lecture 17 – Milankovitch Cycles
- Lecture 18 – Geological Cycles
- Lecture 19 – The Next Steps
- Lecture 20 – Summary/Discussion

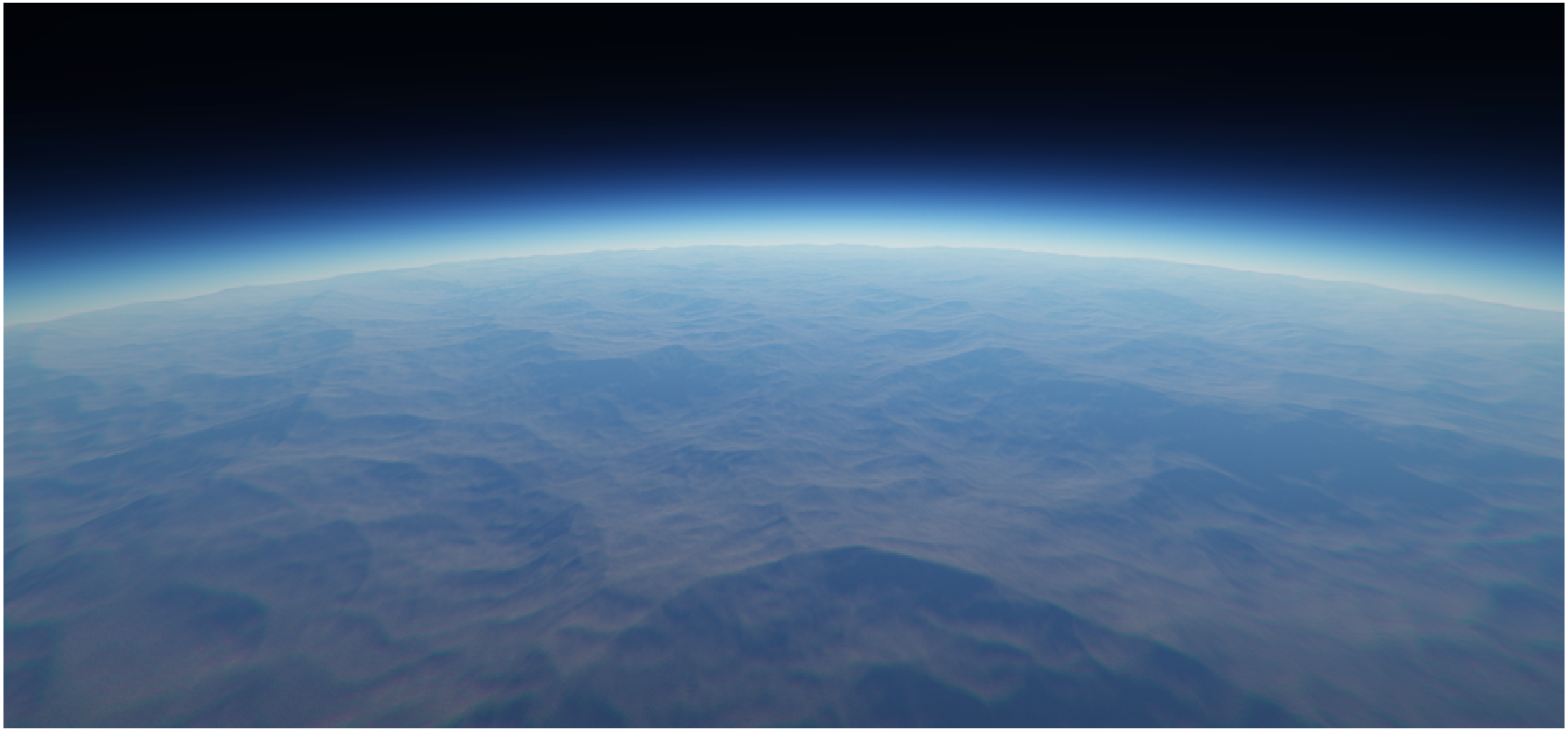


Isotopic Abundance
Birth Environ.
Orbit
Cosmic Rays
Galactic Location
Tides
Orbital Evolution
Impact Flux
Elem. Abundance
Composition & Structure
Luminosity
Magnetic Field
Radius
Orb. Elements
Metallicity
Age
Dust
Minor Planets
Wind
Stellar Effects
Planetary System
Orbits
Rotation
Mass
Sibling Planets
Activity
Companions
Satellites
Spectral Energy Distribution
Liquid Surface Water
Masses
Spectrum
Mass
Obliq.
Radius
Biology
Planetary Properties
Rot. Rate
Dynamics
Albedo
Atm. Dynamics
Technology
Liquids
Oblateness
Clouds
Surface
Outgassing
Surf. Temp.
Comp.
Int. Comp.
Inhomogeneities
UV Shielding
Atmosphere
Magnetic Field
Structure
Interior
Isotopic Ratios
Atm. Escape
Atm. Evol.
Surface Pressure
Density
Internal Evol.
Tectonic Regime
Atm. Structure

Planetary Atmospheres

A key component of studying planetary atmospheres is to develop models of pressure-temperature profiles. Some of the things that affect the profiles are:

- mass, radius, surface gravity
- composition, abundances
- external heat (star)
- internal heat (cooling, radioactivity)
- age



Primary atmosphere

- A planet's *primary atmosphere* comes from nebular material in accretion disk.
  - Mainly H, H_2 and He.
  - Trace elements also present in CO_2, CH_4, N_2, H_2O, NH_3.

- If planet's gravity not strong enough or surface temperature is too large, these elements escape, leaving planet without an atmosphere.

- Solar wind can also drag material from the atmosphere.
  - Relevant for planets without significant magnetospheres (e.g., Mars).

- For the terrestrial planets, most of the H escaped, leaving heavier gases such as argon, neon and ammonia concentrated near the surface.

Secondary atmosphere

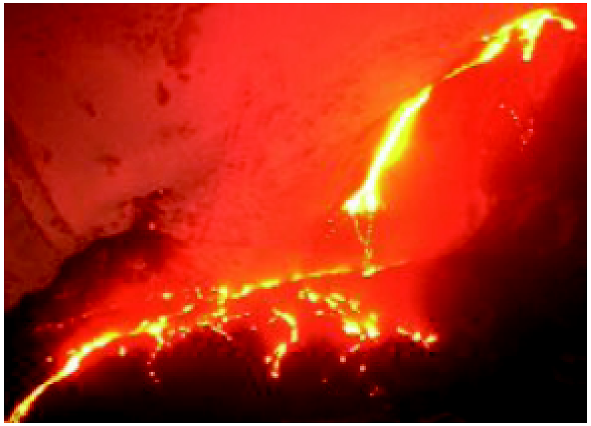
- Rocks and planetesimals which combined to form each planet had trapped gasses.

- During formation, gases released from interior.
  - Differentiation caused them to rise to the outer surface of the planet.
  - Released via volcanism.

- Comets/meteors containing water and gas collided with the planets (H_2O, CH_4, CO_2).

- Volcanic gasses account for most of Earth's atmosphere. Primitive atmosphere contained H_2, H_2O, CO and H_2S.

- Biological activity: photosynthesis converts CO_2 to O_2.



Mount Etna - March 2005
(credit Reuters/Irish Times)

Atmospheric Scale Height

The scale height of an atmosphere is the vertical distance over which the density and pressure fall by a factor of $1/e$. $= 0.368$

Derivation of scale height:

$$dP = -\rho g\,dz = -\left(\frac{\mu m_H P}{kT}\right) g\,dz$$

where $PV = NkT = \dfrac{M}{\mu m_H} kT$

μ = mean molecular weight

m_H = mass of hydrogen atom

P_0 = pressure at $z=0$

$$\int_{P_0}^{P} \frac{dP}{P} = \int_0^z -\left(\frac{\mu m_H g}{kT}\right) dz$$

$$\ln P \Big|_{P_0}^{P} = -\frac{\mu m_H g}{kT} z \Big|_0^z$$

$$\ln P - \ln P_0 = -\frac{\mu m_H g}{kT} z$$

$$\ln \frac{P}{P_0} = -\frac{\mu m_H g}{kT} z$$

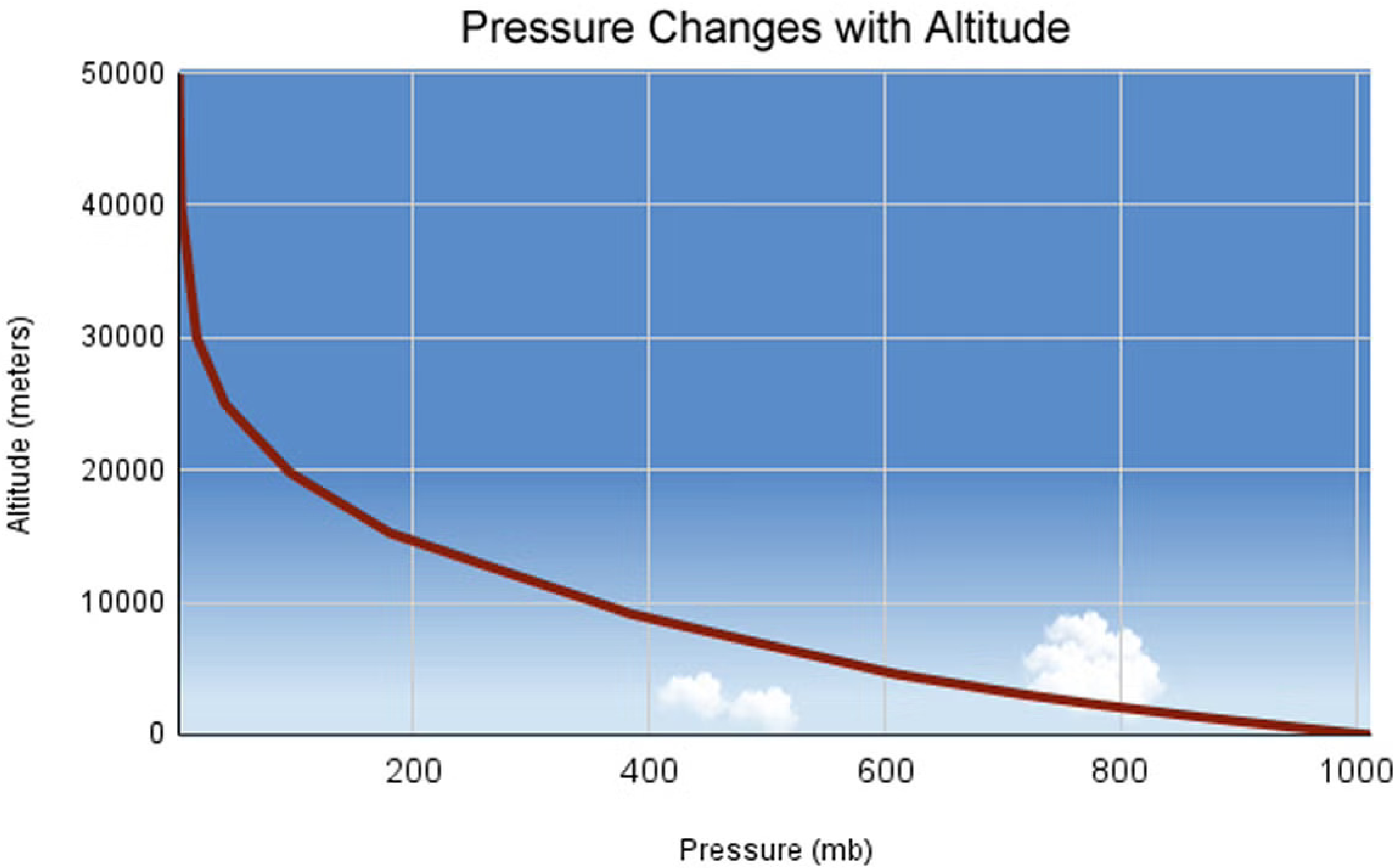
$$\Rightarrow \quad P = P_0 e^{-\frac{\mu m_H g}{kT} z}$$

$$= P_0 e^{-z/H}$$

where scale height H is given by

$$H = \frac{kT}{\mu m_H g}$$

For Earth, $H \sim 8$ km.


Pressure Changes with Altitude
Altitude (meters)
50000
40000
30000
20000
10000
0
200
400
600
800
1000
Pressure (mb)

Thermal Structure

1. Planetary Equilibrium Temperature

We can estimate the temperature of a planet by treating it as a blackbody and equating flux absorbed with flux emitted.

$$F_{in} = F_{out}$$

$$F(1-A_B)\pi R_p^2 = 4\pi R_p^2 \sigma T_p^4 \qquad F = \frac{L_*}{4\pi a^2}$$

$$T_p = \left[\frac{(1-A_B)F}{4\sigma}\right]^{1/4} \qquad A_B = \text{Bond albedo}$$

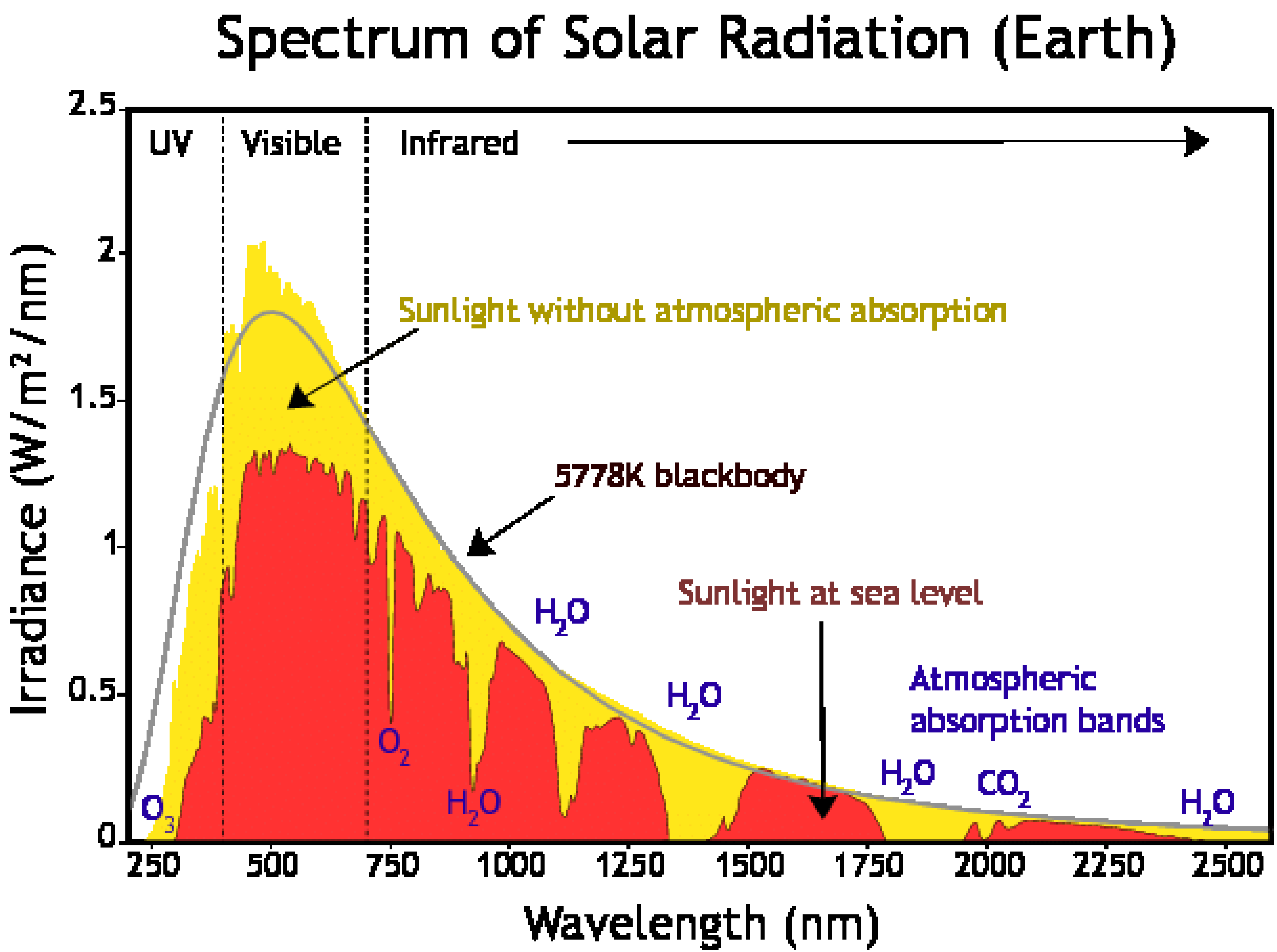
This assumes the flux is reradiated over entire surface area. In other words, assumes 100% heat redistribution. That will depend on atmosphere, rotation rate, wind speeds, etc.

Table E.10 **Basic Atmospheric Parameters for Venus, Earth, Mars and Titan**[a]

Parameter	Venus	Earth	Mars	Titan
Mean heliocentric distance (AU)	0.723	1.000	1.524	9.543
Geometric albedo A_0	0.84	0.367	0.15	0.21
Bond albedo	0.75	0.306	0.25	0.20
Surface temperature (K)	737	288	215	93.7
Equilibrium temperature (K)	232	255	210	85
Exobase temperature (K)	270–320	800–1250	200–300	149
Surface pressure (bar)	92	1.013	0.00636	1.47
Scale height at surface (km)	16	8.5	11	20
Adiabatic lapse rate (K/km)	10.4	9.8	4.4	1.4

[a] All values from Table 4.2 in de Pater and Lissauer (2010). References are provided therein.

Consider the atmospheric pressure at the summit of Mount Everest (elevation of 8.8 km above sea level). The Earth's atmospheric scale height is 8 km. Calculate the altitude needed to reach the same atmospheric pressure on a super-Earth with $a = 0.1$ AU, $M_p = 10M_\oplus$, $R_p = 1.5R_\oplus$, and is orbiting a solar analog. Assume the surface pressure on the planet is the same as the surface pressure on Earth.


Spectrum of Solar Radiation (Earth)
UV
Visible
Infrared
Sunlight without atmospheric absorption
5778K blackbody
Sunlight at sea level
Atmospheric absorption bands
O_3
O_2
H_2O
H_2O
H_2O
H_2O
CO_2
H_2O
Irradiance (W/m²/nm)
Wavelength (nm)
2.5
2
1.5
1
0.5
0
250
500
750
1000
1250
1500
1750
2000
2250
2500

2. Radiative Transfer

The change in intensity, dI_λ, of a ray of wavelength λ as it travels through a gas is proportional to its intensity, I_λ, the distance traveled, ds, and the density of the gas, ρ.

$$dI_\lambda = -\alpha_\lambda \rho I_\lambda \, ds \qquad \ldots (1)$$

α_λ is called the absorption (or extinction) coefficient, or opacity. The minus sign shows the intensity decreases with distance.

An increase in intensity due to emission is given by

$$dI_\lambda = j_\lambda \rho \, ds \qquad \ldots (2)$$

where j_λ is the emission coefficient. Combining (1) and (2) gives

$$dI_\lambda = -\alpha_\lambda \rho I_\lambda \, ds + j_\lambda \rho \, ds$$

Rearranging:

$$-\frac{1}{\alpha_\lambda \rho} \frac{dI_\lambda}{ds} = I_\lambda - \frac{j_\lambda}{\alpha_\lambda}$$

$$= I_\lambda - S_\lambda$$

where $S_\lambda \equiv \frac{j_\lambda}{\alpha_\lambda}$ is the source function.

This is called the equation of radiative transfer.

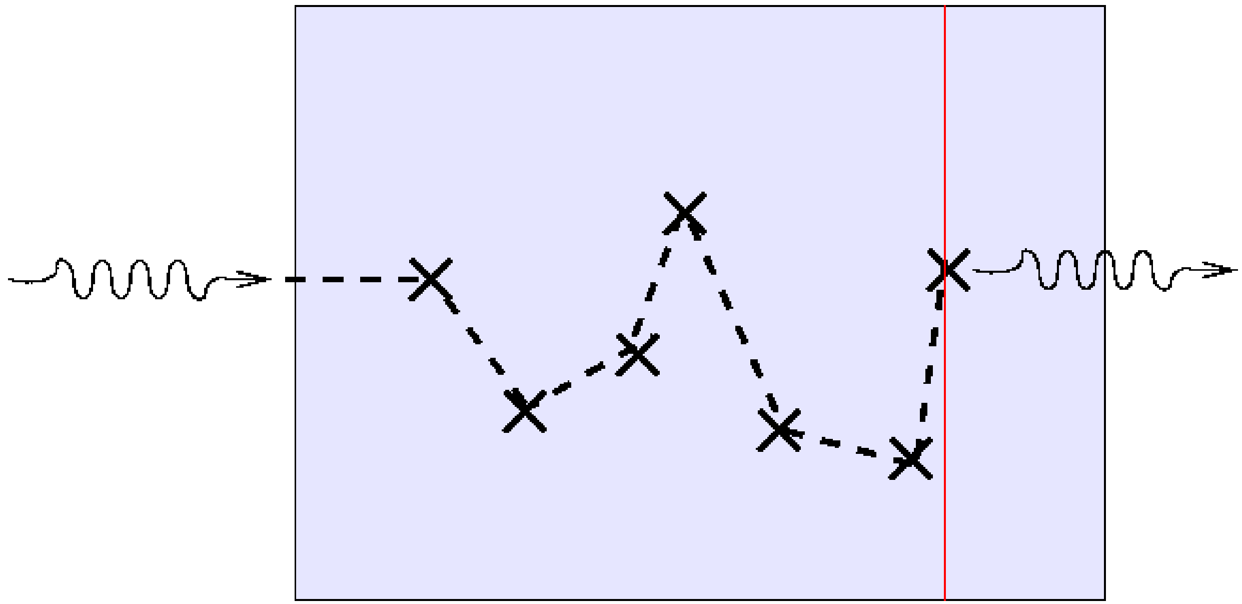
We define the optical depth, τ_λ, as

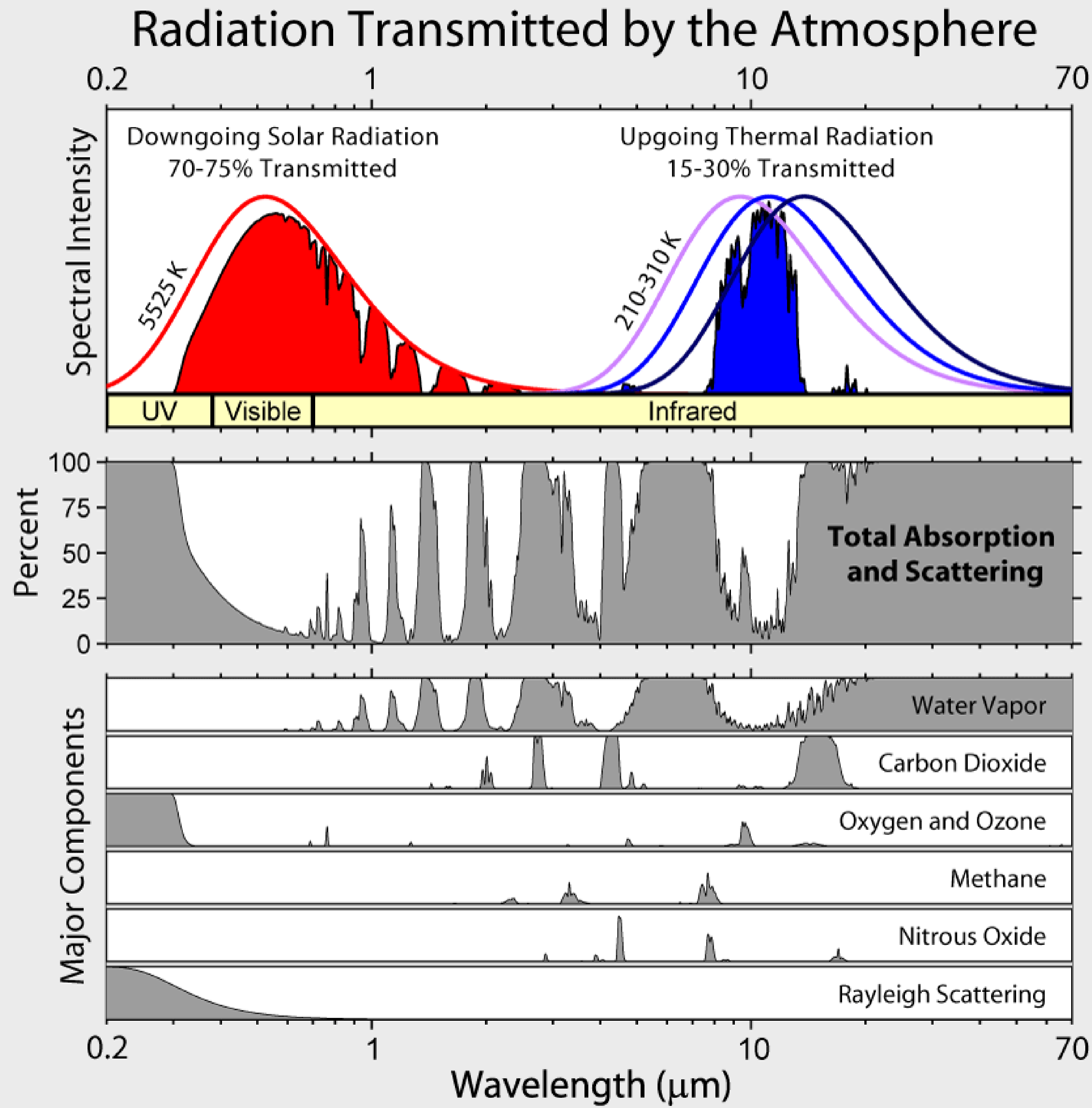
$$d\tau_\lambda = -\alpha_\lambda \rho \, ds$$

and so is the integral over the extinction coefficient:

$$\tau_\lambda \equiv \int_{s_1}^{s_2} \alpha_\lambda(s)\, \rho(s)\, ds$$

If the optical depth is $\tau_\lambda = 1$, the intensity of the ray along its path will decline by a factor of e^{-1}. The optical depth may be thought of as the number of mean free paths over the distance ds along the ray's path. We typically see no deeper into an atmosphere at a given wavelength than $\tau_\lambda \approx 1$. If $\tau_\lambda \gg 1$, the gas is said to be optically thick; if $\tau_\lambda \ll 1$, the gas is optically thin.

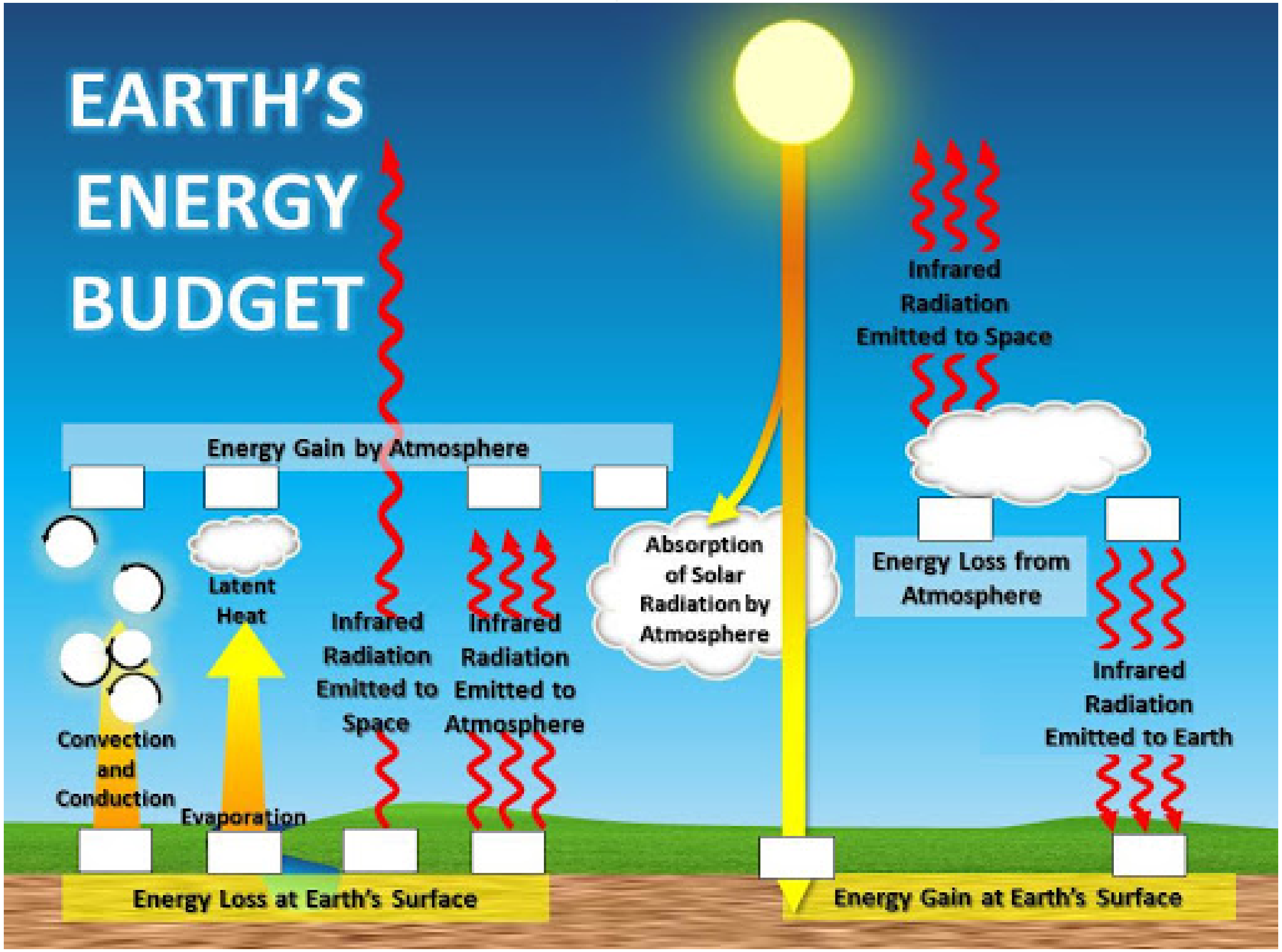




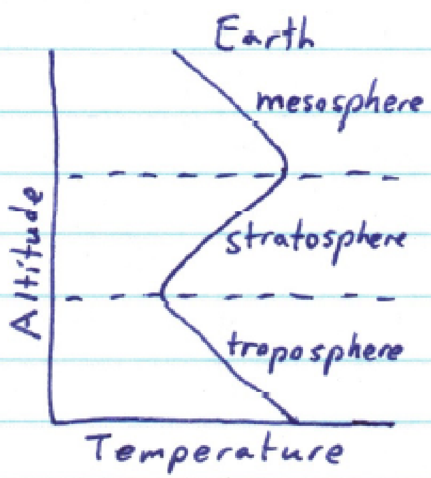
Radiation Transmitted by the Atmosphere
0.2
1
10
70
Spectral Intensity
Downgoing Solar Radiation
70-75% Transmitted
Upgoing Thermal Radiation
15-30% Transmitted
5525 K
210-310 K
UV
Visible
Infrared
Percent
100
75
50
25
0
Total Absorption
and Scattering
Major Components
Water Vapor
Carbon Dioxide
Oxygen and Ozone
Methane
Nitrous Oxide
Rayleigh Scattering
Wavelength (μm)

3. Processes that affect temperature

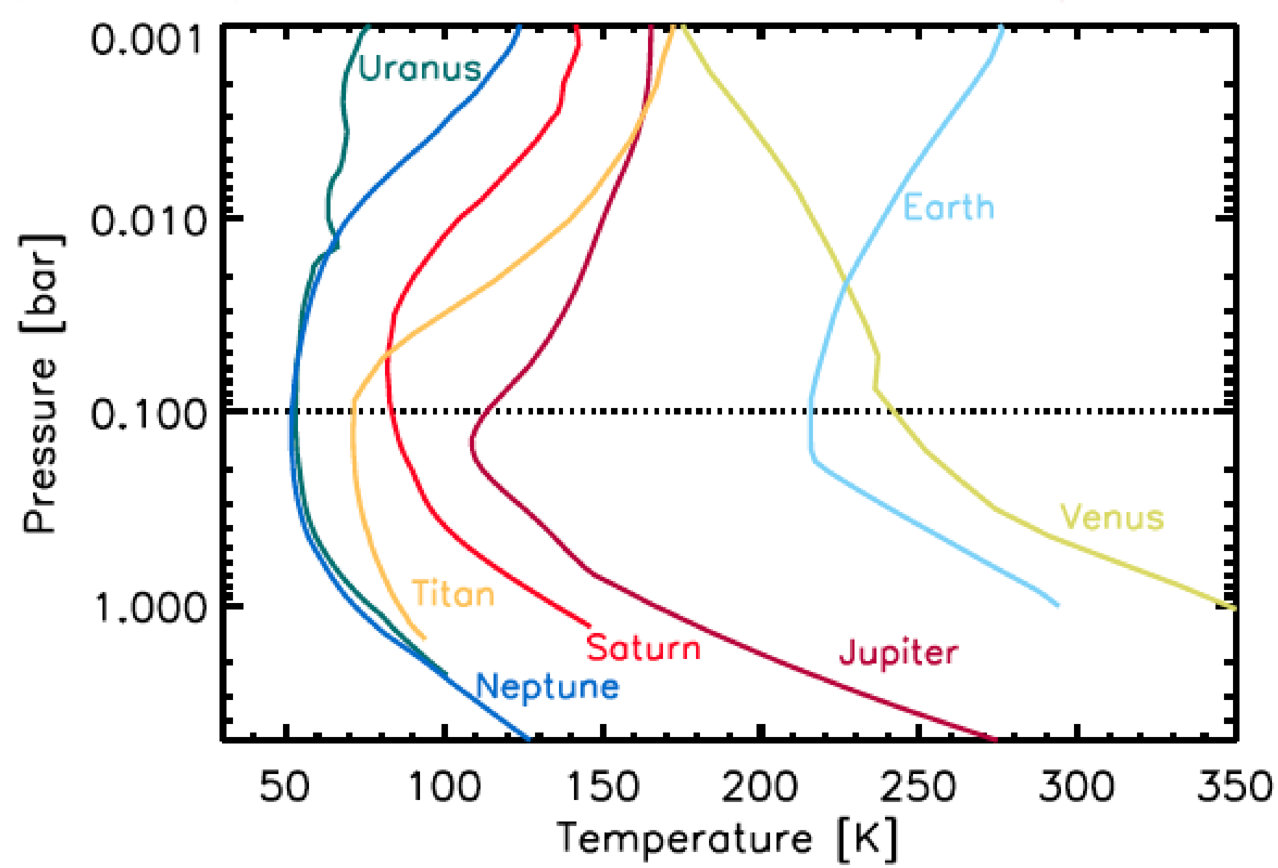
- Irradiation at the top of the atmosphere (star).
- Energy from internal heat sources.
- Chemical reactions in the atmosphere (changes opacity).
- Clouds/haze (change opacity and local temperature).
- Volcanic outgassing
- Chemical interactions between atmosphere and surface.
- Biochemical processes.



4. Inversion layer



Usually temperature decreases as altitude increases since the atmosphere is heated from below because the ground reradiates energy received from the star. Often a planetary atmosphere will have a layer in which energy is released slower than it is absorbed, resulting in a temperature inversion. These layers can yield clues t cloud compositions.



Atmospheric Composition

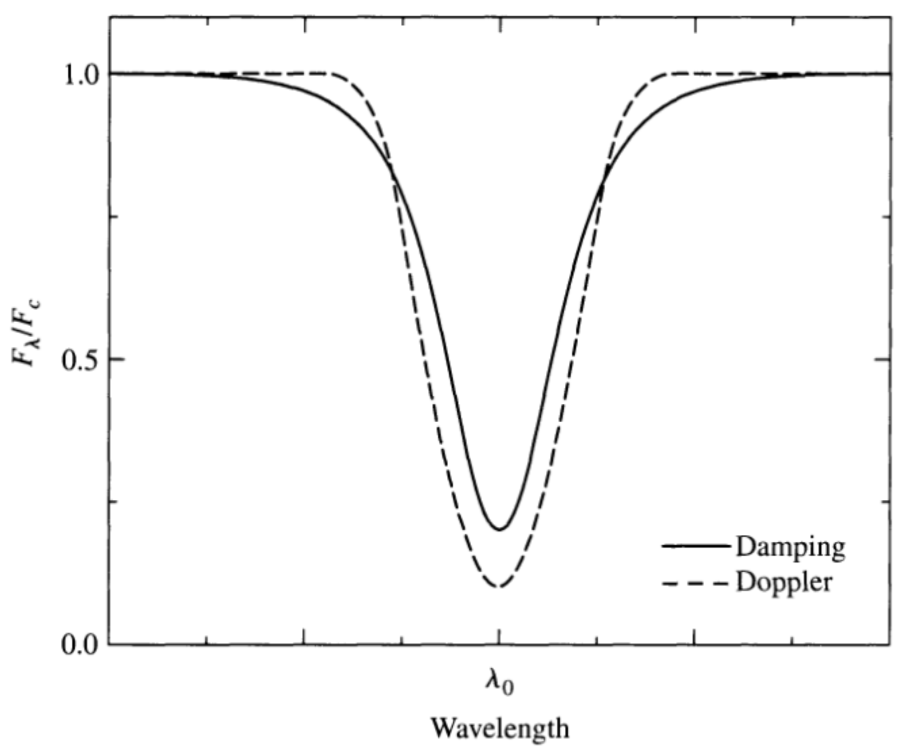
We can learn directly about atmospheric composition from absorption lines resulting from starlight passing through the atmosphere during transit.

Absorption lines are fundamentally caused by the extinction and emission coefficients of the atmospheric gas along the path of the ray. The width of the line profiles is determined by:

(a) Natural Damping: Lorentz profile

(b) Doppler Broadening: Voigt profile

(c) Pressure or Collisional Broadening



Atmospheric escape

- ***Thermal escape*** is when heating of an atmosphere allows molecules to escape. Examples include Jeans’ escape and Hydrodynamic escape.

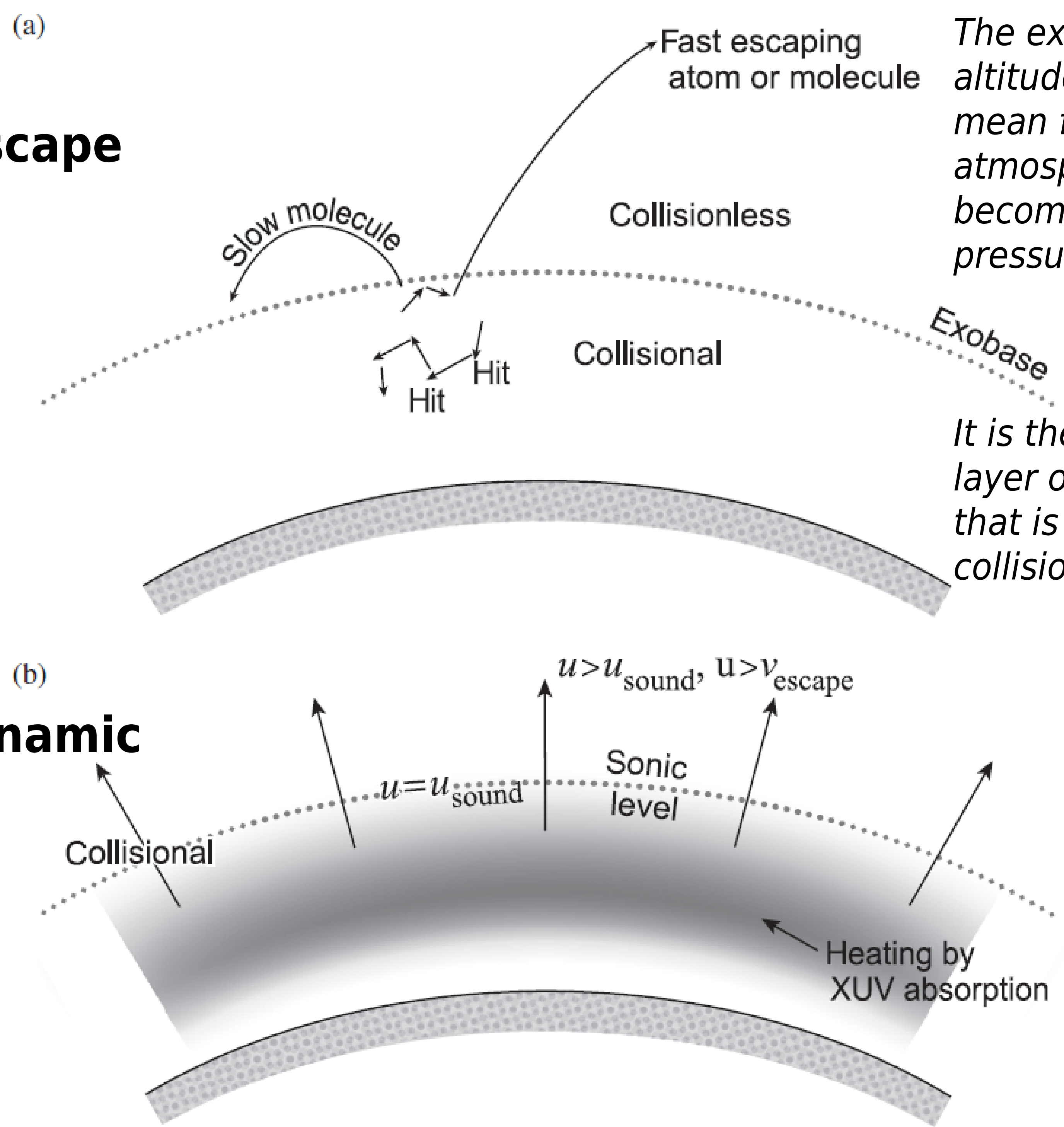
- ***Suprathermal (or non-thermal) escape*** refers to loss processes that affect either neutral species or ions that attain a velocity significantly greater than that corresponding to the background neutral temperature. Examples include Photochemical escape, Charge exchange, Ion pickup, Sputtering, The polar wind, and Bulk removal.

- ***Impact erosion*** occurs when the hot vapor plume or high-speed ejecta associated with a large asteroid or comet impact imparts sufficient kinetic energy to atmospheric molecules for them to escape en masse.

Jeans' escape

The exobase is the altitude where the mean free path of atmospheric particles becomes equal to one pressure scale height.

It is the uppermost layer of an atmosphere that is essentially collisionless.



Hydrodynamic escape

Atmospheric Loss

Consider a planet with escape velocity

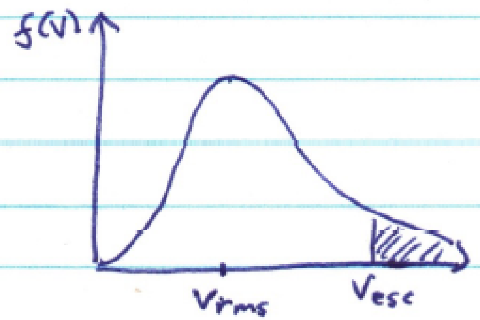
$$V_{esc} = \sqrt{\frac{2GM_p}{R_p}}$$

The rms velocity of molecules with mass m at the surface is given by

$$V_{rms} = \sqrt{\frac{3kT}{m}}$$

For a thermosphere of T ~1000 K
v = 5 km/s < 10.8 km/s at the exobase

So atmospheric escape will occur when $V_{rms} > V_{esc}$. HOWEVER, this assumes a single mean free path to escape.

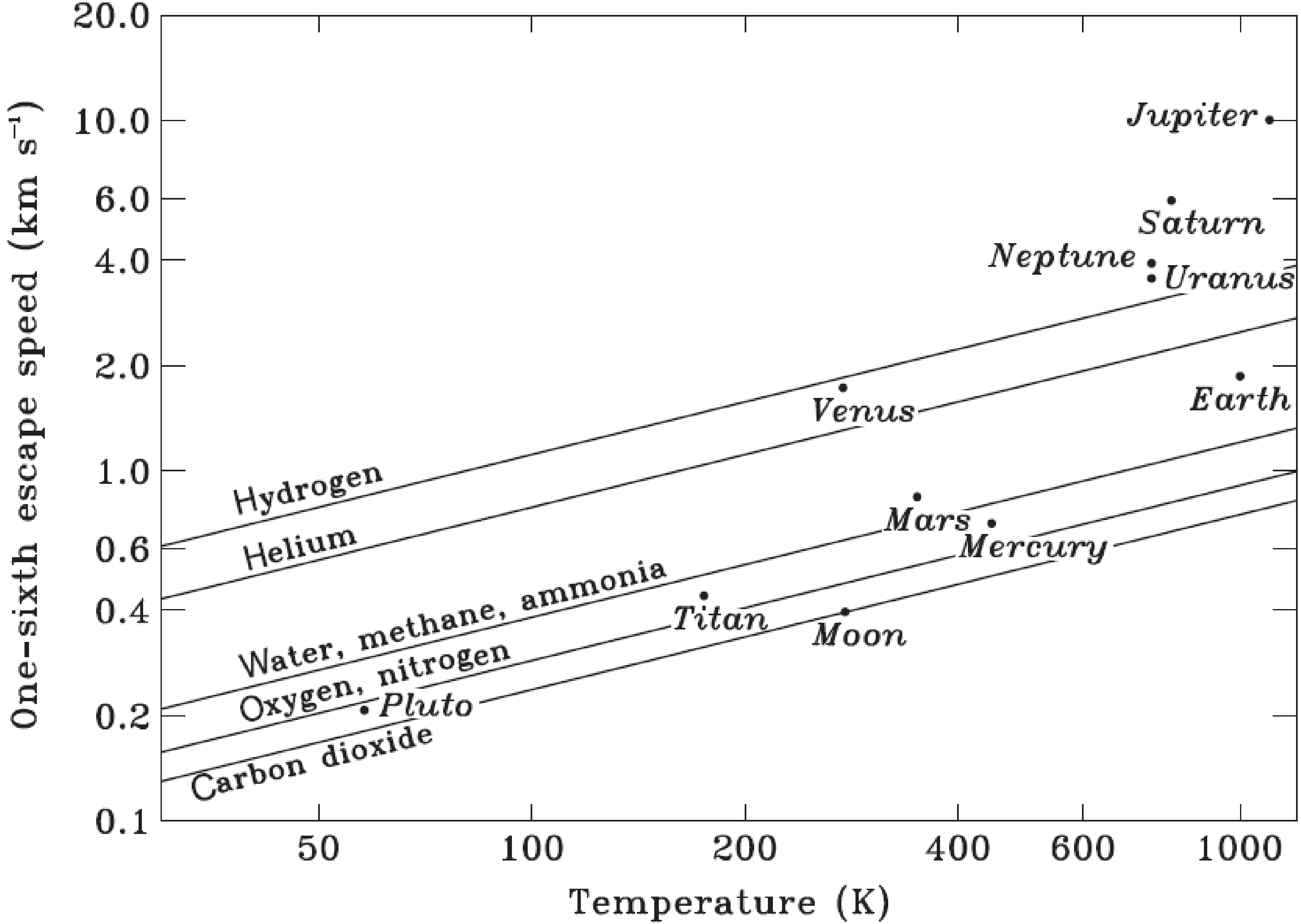


Maxwell–Boltzmann velocity distribution

Because of collisions, atmospheric escape only occurs at the tail of the velocity distribution. Significant escape then requires $6\, V_{rms} > V_{esc}$.

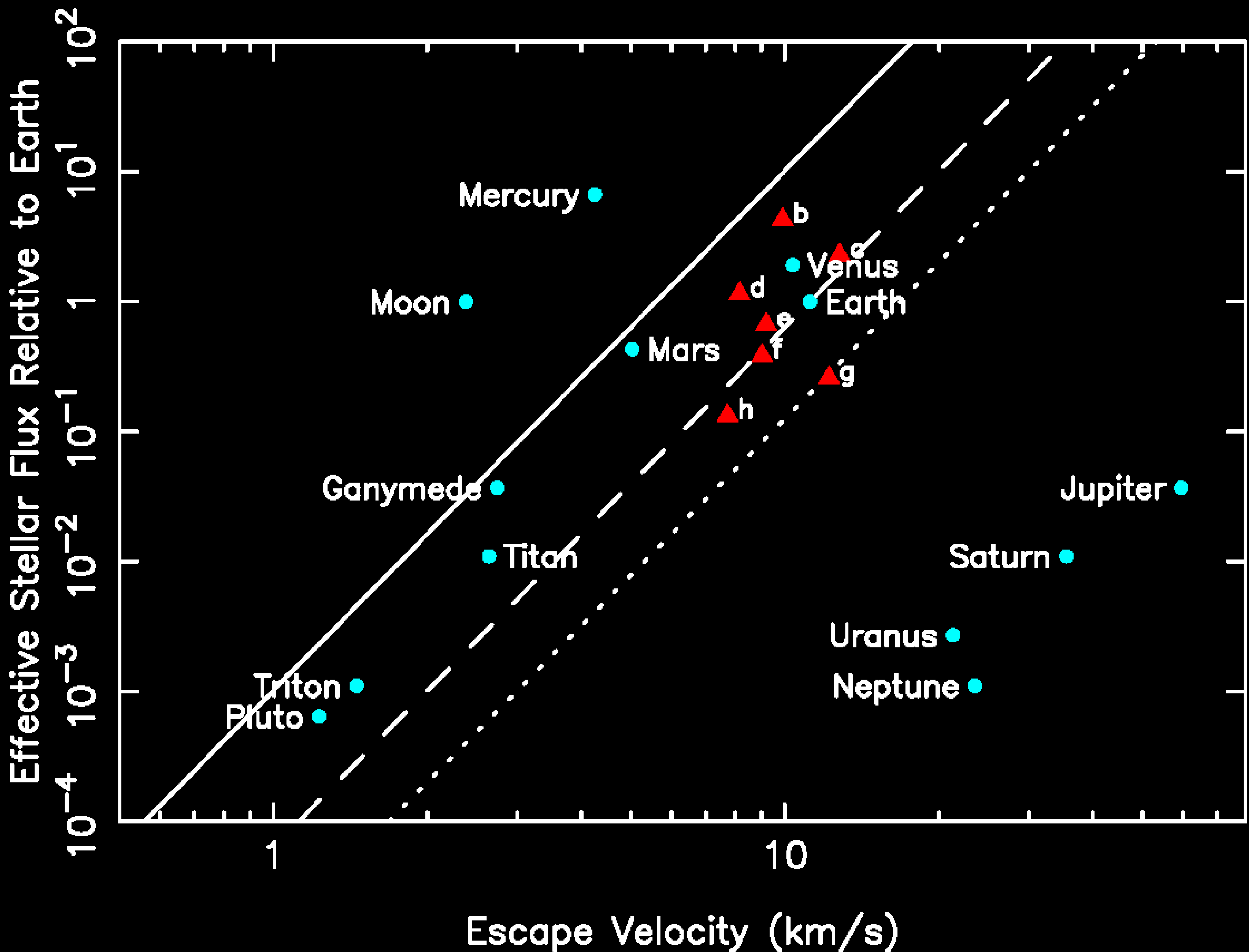
$$6\sqrt{\frac{3kT}{m}} > \sqrt{\frac{2GM_p}{R_p}}$$

$$\Rightarrow T > \frac{1}{54}\,\frac{GM_p}{R_p}\,\frac{m}{k}$$



20.0
10.0
6.0
4.0
2.0
1.0
0.6
0.4
0.2
0.1
One-sixth escape speed (km s^{-1})
50
100
200
400
600
1000
Temperature (K)
Jupiter
Saturn
Neptune
Uranus
Earth
Venus
Mars
Mercury
Titan
Moon
Pluto
Hydrogen
Helium
Water, methane, ammonia
Oxygen, nitrogen
Carbon dioxide

The Cosmic Shoreline



Zahnle & Catling, 2017, ApJ, 843, 122
Roettenbacher & Kane, 2017, ApJ, 851, 77