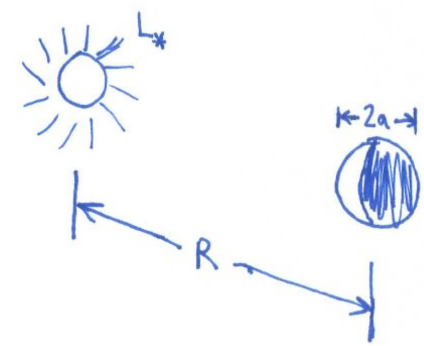
Planetary Temperatures



- How does Sunlight heat a planet with no atmosphere? This is similar to our dust grain heating problem
- **First pass**: Consider a planet of radius *a* at a distance *R* from a star of luminosity L_*. The energy received by the planet per time per area is,

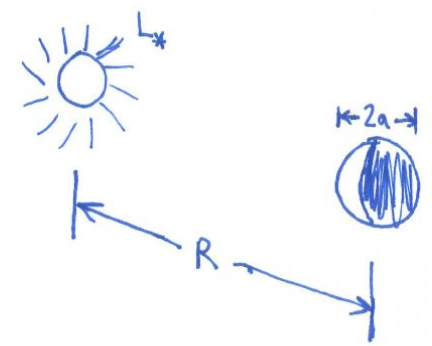
$$\frac{L_*}{4\pi R^2}\pi a^2$$

- The planet is heated to a temperature, *T*, & re-emits radiation in thermal infrared. The amount of radiation released is

$$4\pi a^2 \sigma T^4$$

Planetary Temperatures

- By equating both of these expressions,



$$\frac{L_*}{4\pi R^2}\pi a^2 = 4\pi a^2 \sigma T^4$$

- And solving for *T*,

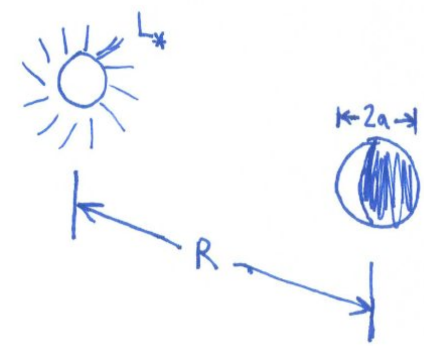
$$T = \frac{1}{2}\left(\frac{L_*}{\pi R^2 \sigma}\right)^{1/4}$$

- So, plugging in the appropriate numbers for Earth,

$$T = \frac{1}{2}\left(\frac{L_*}{\pi R^2 \sigma}\right)^{1/4} = \frac{1}{2}\left[\frac{3.83 \times 10^{26}\ \mathrm{W}}{\pi (1.5 \times 10^{11}\ \mathrm{m})^2 \times 5.67 \times 10^{-8}\ \mathrm{W\ m^{-2}\ K^{-4}}}\right]^{1/4} = 278\ \mathrm{K}$$

Planetary Temperatures

- **Second pass:** But, of course, planets reflect a certain fraction of the light that strikes then. The **albedo**, *A*, is defined as the fraction of light incident on an object that is reflected back into space. So, in actuality, the energy absorbed by the planet time^{-1} area^{-1} is,



$$\frac{L_*}{4\pi R^2}\pi a^2(1-A)$$

- And thus, the temperature expression becomes,

$$T = \frac{1}{2}\left(\frac{L_*[1-A]}{\pi R^2 \sigma}\right)^{1/4}$$

Planetary Temperatures

- In terms of albedo, the general breakdown is -

  Albedo = 0 (no reflection)
  Albedo = 1 (all reflected)
  Albedo = 0.7 (cloud, snow, ice)
  Albedo = 0.1-0.25 (rocks)

Planetary Temperatures

- The Earth has an albedo across all wavelengths (I.e., a **Bond** albedo) of 0.306. So, the *T* of the Earth is,

$$T = \frac{1}{2}\left(\frac{L_*(1-A)}{\pi R^2 \sigma}\right)^{1/4} = \frac{1}{2}\left[\frac{3.83 \times 10^{26}\ \mathrm{W} \times (1-0.306)}{\pi(1.5 \times 10^{11}\ \mathrm{m})^2 \times 5.67 \times 10^{-8}\ \mathrm{W\ m^{-2}\ K^{-4}}}\right]^{1/4} = 253\ \mathrm{K}$$

Table X.1. *The visual and Bond albedoes of the planets.*

Planet	Visual albedo	Bond albedo
Mercury	0.106	0.119
Venus	0.650	0.750
Earth	0.367	0.306 ←
Mars	0.150	0.250
Jupiter	0.520	0.343
Saturn	0.470	0.342
Uranus	0.510	0.300
Neptune	0.410	0.290
Pluto	0.300	0.367

Planetary Temperatures

- The table below shows the results of the calculation for the terrestrial planets & the moon -

Table 10.2 Temperatures of the Terrestrial Worlds

World	Distance to Sun (AU)	Albedo (0 = black, 1 = white)	Length of Day (Earth days)	"No Greenhouse" Temperature*	Observed Average Surface Temperature
Mercury	0.38	0.11	176	440 K	700 K (day), 100 K (night)
Venus	0.72	0.72	117	230 K	740 K
Earth	1.00	0.36	1	250 K	288 K
Moon	1.00	0.07	28	273 K	400 K (day), 100 K (night)
Mars	1.52	0.25	≈1	218 K	223 K

*Assumes rapid rotation, in which case day and night temperatures would be the same.

$$T = \frac{1}{2}\left(\frac{L_*(1-A)}{\pi R^2 \sigma}\right)^{1/4} = \frac{1}{2}\left[\frac{3.83\times10^{26}\ \mathrm{W}\times(1-0.306)}{\pi(1.5\times10^{11}\ \mathrm{m})^2\times5.67\times10^{-8}\ \mathrm{W\ m^{-2}\ K^{-4}}}\right]^{1/4} = 253\ \mathrm{K}$$

Planetary Temperatures

- We have, of course, ignored a couple of things:
- First, planets rotate, & the speed of the rotation affects whether the planet has a nearly uniform temperature, or if it has a “day” & “night”-side temperatures. For no, rotation, only half of the planet is illuminated, thus,

$$\frac{L_*(1-A)}{4\pi R^2}\pi a^2 = \mathbf{2}\pi a^2 \sigma T^4 \quad \rightarrow \quad T = \left[\frac{L_*(1-A)}{\mathbf{8}\pi\sigma R^2}\right]^{1/4}$$

Table 10.2 Temperatures of the Terrestrial Worlds

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Mars	1.52	0.25	≈1	218 K	223 K

*Assumes rapid rotation, in which case day and night temperatures would be the same.

Planetary Temperatures

- For a non-rotating Mercury,

$$T = \left(\frac{L_*(1-A)}{8\pi R^2 \sigma}\right)^{1/4} = \frac{1}{2}\left[\frac{3.83 \times 10^{26}\ \mathrm{W} \times (1 - 0.119)}{8\pi(0.38\ \mathrm{AU} \times 1.5 \times 10^{11}\ \mathrm{m\ AU^{-1}})^2 \times 5.67 \times 10^{-8}\ \mathrm{W\ m^{-2}\ K^{-4}}}\right]^{1/4} = 520\ \mathrm{K}$$

- I.e., lower than 700 K, but higher by $2^{1/4}$ than the rapid rotation approximation.

Table 10.2 Temperatures of the Terrestrial Worlds

World	Distance to Sun (AU)	Albedo (0 = black, 1 = white)	Length of Day (Earth days)	"No Greenhouse" Temperature*	Observed Average Surface Temperature
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*Assumes rapid rotation, in which case day and night temperatures would be the same.

Planetary Temperatures

- Second, planets have atmospheres, the composition of which will affect the planet's surface temperature. We'll talk about the Greenhouse effect momentarily

Table 10.2 Temperatures of the Terrestrial Worlds

World	Distance to Sun (AU)	Albedo (0 = black, 1 = white)	Length of Day (Earth days)	"No Greenhouse" Temperature*	Observed Average Surface Temperature
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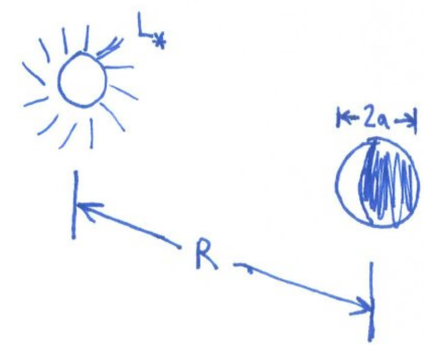
*Assumes rapid rotation, in which case day and night temperatures would be the same.

Planetary Temperatures

- **Third pass:** we can include an emissivity, ε, which is the fraction on incident radiation which is emitted by the planet. Typically, ε ~ 0.8 - 0.9.



$$\frac{L_*}{4\pi R^2}\pi a^2(1-A) = 4\pi a^2 \sigma \epsilon T^4$$

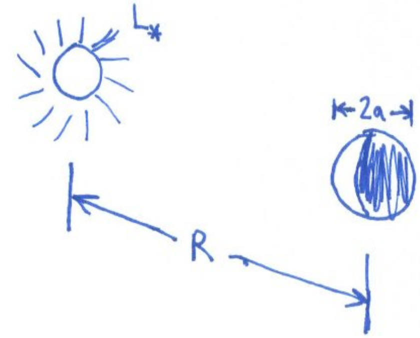
- And thus, the temperature expression becomes,

$$T = \frac{1}{2}\left[\frac{L_*(1-A)}{\sigma\epsilon\pi R^2}\right]^{1/4}$$

- The *T* is higher by a factor of 1.03 - 1.06 because of ε.

Planetary Temperatures

- Note that atmospheres affect *A* & ε because of clouds, surface volatiles (ocean, polar caps), gaseous molecular absorption, & Rayleigh scattering.

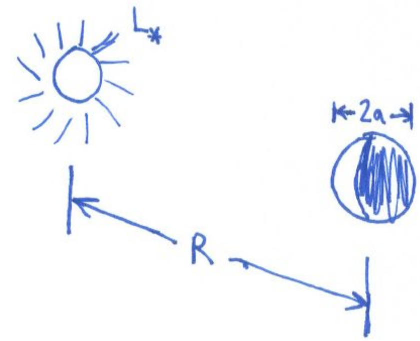


$$T = \frac{1}{2}\left[\frac{L_*(1-A)}{\sigma\epsilon\pi R^2}\right]^{1/4}$$

Planetary Temperatures

- **Fourth pass:** planets can have internal heat sources. If the internal heat source is L_p, then,

$$\frac{L_*}{4\pi R^2}\pi a^2(1-A) + L_p = 4\pi a^2 \sigma \epsilon T^4$$



- Observations of the giant planets show that L_p is significant.
- But where does L_p come from? Maybe heat of formation. Consider,

$$t_K = \frac{\text{total gravitational binding energy}}{\text{rate of energy loss}} \sim \frac{GM^2}{R}\frac{1}{L}$$

Planetary Temperatures

$$t_K = \frac{\text{total gravitational binding energy}}{\text{rate of energy loss}} \sim \frac{GM^2}{R}\frac{1}{L}$$

	M (kg)	R(m)	L(W)	t_K (s)	t_K (yr)
Jupiter	2×10^{27}	7×10^{7}	3×10^{17}	1.3×10^{19}	4×10^{11}
Saturn	6×10^{26}	6×10^{7}	8×10^{16}	5×10^{18}	1×10^{11}
Uranus	9×10^{25}	2×10^{7}	$< 2 \times 10^{15}$	$> 1.3 \times 10^{19}$	$> 4 \times 10^{11}$
Neptune	1×10^{26}	2×10^{7}	2×10^{15}	1.7×10^{19}	6×10^{11}

- For t_K > age of planets, the observed heat loss may be heat of formation
- For Saturn, t_K < age of Saturn. Detailed calculations verify this.

Planetary Temperatures

	M (kg)	R(m)	L(W)	t_K (s)	t_K (yr)
Jupiter	2×10^{27}	7×10^{7}	3×10^{17}	1.3×10^{19}	4×10^{11}
Saturn	6×10^{26}	6×10^{7}	8×10^{16}	5×10^{18}	1×10^{11}
Uranus	9×10^{25}	2×10^{7}	$<2 \times 10^{15}$	$>1.3 \times 10^{19}$	$>4 \times 10^{11}$
Neptune	1×10^{26}	2×10^{7}	2×10^{15}	1.7×10^{19}	6×10^{11}

- For Saturn, **helium rain** is responsible for heat loss. At low temperature and pressure, liquid helium does not dissolve with liquid hydrogen. A deficit of helium is observed in the outer atmosphere of Saturn.
- Energy is produced through friction with atmosphere

Atmospheres: Pressure vs. Height

- Consider again the equation of hydrostatic equilibrium

$$\frac{dp}{dh} = -\rho(h)g(h)$$

- Where *g(h)* is the acceleration due to gravity & $\rho(h)$ is the density. Consider also the ideal gas law

$$p = \frac{\rho k T}{\mu m_H}$$

- Assume further that *g(h)* = *g*, *T* = constant, & μ = constant.

$$\frac{dp}{dh} = -\left(\frac{p\mu m_H}{kT}\right)g$$

Atmospheres: Pressure vs. Height

- Taking

$$\frac{dp}{dh} = -\left(\frac{p\mu m_H}{kT}\right) g$$

- and integrating both sides,

$$\int_{P(0)}^{P} \frac{dp}{p} = -\left(\frac{\mu m_H g}{kT}\right) \int_{h=0}^{h} dh$$

$$\ln\left[\frac{P}{P(0)}\right] = -\left(\frac{\mu m_H g}{kT}\right) h$$

$$P(h) = P(0) e^{-\left(\frac{\mu m_H g}{kT}\right) h} = P(0) e^{-\left(\frac{h}{H}\right)}$$

- The quantity *H* is called the scale height.

Atmospheres: Pressure vs. Height

- For the Earth,

$$H = \frac{kT}{\mu m_H g} = \frac{1.38 \times 10^{23} \;\; \mathrm{J\ K^{-1}} \times 290\ \mathrm{K}}{29 \times 1.67 \times 10^{27}\ \mathrm{kg} \times 9.8\ \mathrm{m\ s^{-2}}} \sim 8.4 \times 10^3 \mathrm{m}$$

- or,

$$8.4 \times 10^3 \mathrm{m} \left(\frac{R_{\mathrm{earth}}}{6.4 \times 10^6 \mathrm{m}} \right) = 0.0013 R_{\mathrm{earth}}$$

- I.e., $H << R_{\mathrm{planet}}$

Mass of Atmosphere

- Because the atmosphere is so thin, the mass of the atmosphere can be estimated via,

$$P = \frac{mg}{\text{area}} = \frac{mg}{4\pi R^2}$$

- Rearranging the terms & plugging in numbers for the Earth,

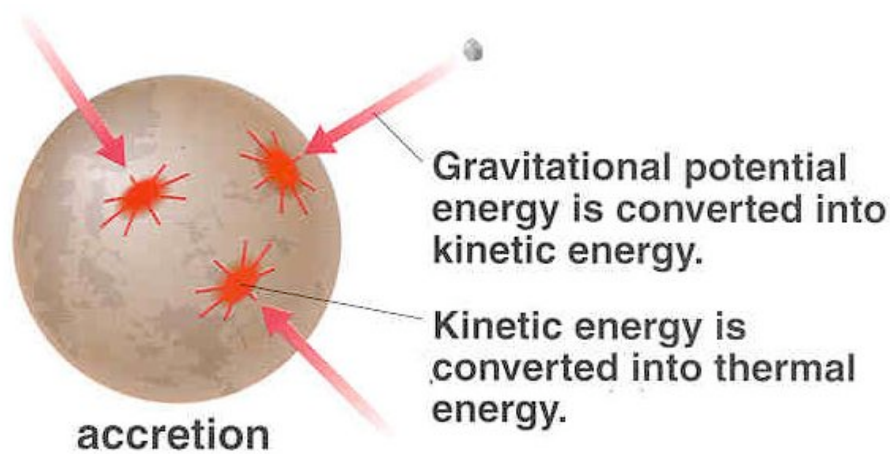
$$m = \frac{4\pi R^2 P}{g} = \frac{4\pi \times (6.4 \times 10^6\ \text{m})^2 \times 10^5\ \text{N m}^{-2}}{9.8\ \text{m s}^{-2}} \sim 5 \times 10^{18}\ \text{kg}$$

- Or,

$$5 \times 10^{18}\ \text{kg} \left(\frac{M_{\text{earth}}}{6.0 \times 10^{24} \text{kg}} \right) = 9 \times 10^{-7} M_{\text{earth}}$$

How does a planet obtain an atmosphere?

- **Capture/Primordial**: it forms with one. This is applicable to the gas giants
- **Outgassing**: it produces one from the material in which the planet is made. This is applicable to the terrestrial planets. They produced their atmosphere as the result of volcanism.



How does a planet hold an atmosphere?

- First, define the **exosphere** as the height, h_{atm}, at which the atmosphere is so thin that a gas molecular has a mean-free-path of infinity.
- For the Earth, h_{atm} ~ 500 km, which is,

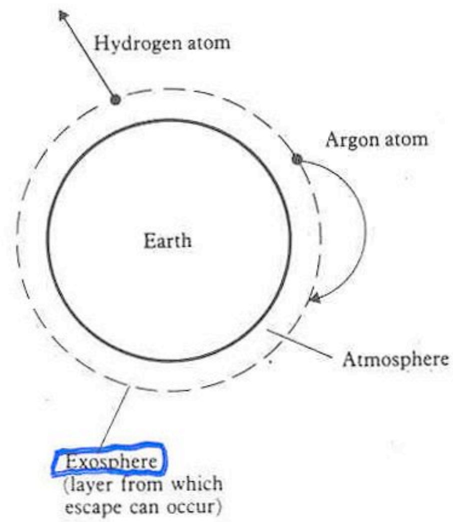


FIGURE 3.3 The light hydrogen atoms (atomic mass = 1) move much faster than the heavier argon atoms (atomic mass = 40) at the same temperature. Thus hydrogen can escape from the Earth, while argon cannot.

$$500 \times 10^3 \text{ m} \left(\frac{R_{\text{earth}}}{6.4 \times 10^6 \text{ m}} \right) = 0.078 R_{\text{earth}}$$

How does a planet hold an atmosphere?

- It must be massive enough…

$$v_{\rm esc} = \sqrt{\frac{2GM_{\rm planet}}{R_{\rm planet} + h_{\rm atm}}} \approx \sqrt{\frac{2GM_{\rm planet}}{R_{\rm planet}}}$$

- And cool enough…

$$kT_{\rm atm} \sim \mu m_H v_{\rm gas}^2$$

$$v_{\rm gas} \sim \sqrt{\frac{kT_{\rm atm}}{\mu m_H}}$$

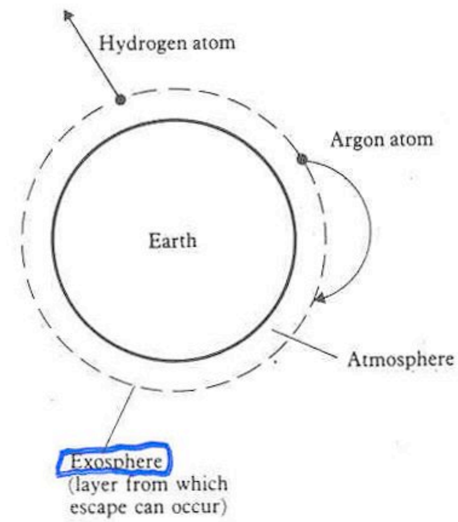


FIGURE 3.3 The light hydrogen atoms (atomic mass = 1) move much faster than the heavier argon atoms (atomic mass = 40) at the same temperature. Thus hydrogen can escape from the Earth, while argon cannot.

- … or the gas will escape.
- For a fixed *T*, lighter atoms escape more readily than heavier atoms because they have higher velocities

How does a planet hold an atmosphere?

- For a gas of mass μm_H in equilibrium at T, the probability distribution of speeds is described by th Maxwell-Boltzmann distribution.

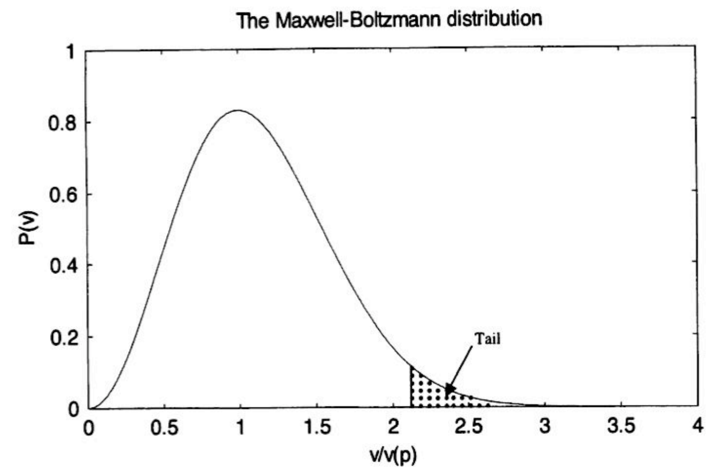


Figure O.4. *The Maxwell–Boltzmann distribution showing the tail which extends to infinity. The velocity is made non-dimensional by dividing by* $v(p) = \sqrt{2kT/\mu}$.

$$p(v) = \left(\frac{2\mu^3 m_H^3}{\pi k^3 T^3}\right)^{1/2} v^2 e^{-\frac{\mu m_H v^2}{2kT}}$$

- Given this, the proportion of molecules with speeds exceeding the escape velocity at any time is

$$f = \int_{v_{\rm esc}}^{\infty} p(v)\,dv$$

Gas Retention

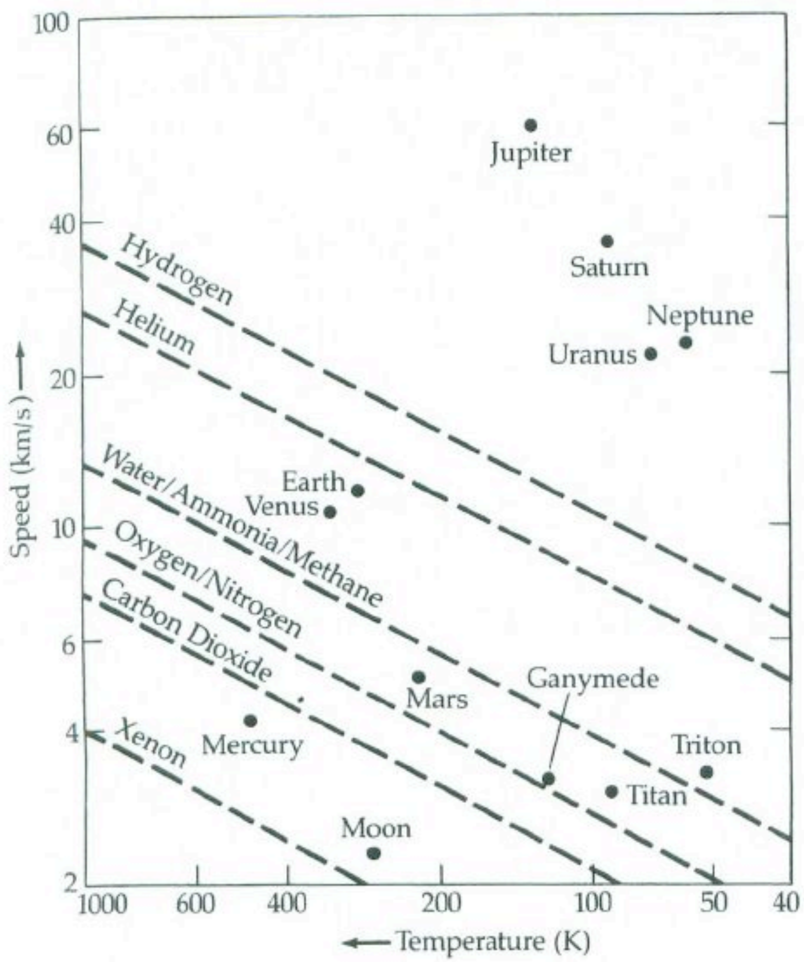


Figure 2–8 Retention of atmospheric gases. Mean molecular speeds are given as a function of temperature, along with the escape speeds for the indicated bodies. The dashed lines show ten times the mean molecular speeds, which defines an essentially infinite lifetime for that component in the atmosphere.

Photo	Planet	Average Distance from Sun (AU)	Temperature†	Relative Size	Average Radius (km)	Average Density (g/cm^3)	Composition	Moons	Rings?
	Mercury	0.387	700 K	·	2,440	5.43	Rocks, metals	0	No
	Venus	0.723	740 K	•	6,051	5.24	Rocks, metals	0	No
	Earth	1.00	290 K	•	6,378	5.52	Rocks, metals	1	No
	Mars	1.52	240 K	·	3,397	3.93	Rocks, metals	2 (tiny)	No
	Most asteroids	2–3	170 K	·	≤500	1.5–3	Rocks, metals	?	No
	Jupiter	5.20	125 K	●	71,492	1.33	H, He, hydrogen compounds‡	16	Yes
	Saturn	9.53	95 K	●	60,268	0.70	H, He, hydrogen compounds‡	18	Yes
	Uranus	19.2	60 K	●	25,559	1.32	H, He, hydrogen compounds‡	17	Yes
	Neptune	30.1	60 K	●	24,764	1.64	H, He, hydrogen compounds‡	8	Yes
	Pluto	39.5	40 K	·	1,160	2.0	Ices, rock	1	No
	Most comets	0–50,000	a few K§	·	a few km?	<1?	Ices, dust	?	No

*Appendix C gives a more complete list of planetary properties. †Surface temperatures for all objects except Jupiter, Saturn, Uranus, and Nepturne, for which cloud-fog temperatures are listed. ‡Includes water (H_2O), methane (CH_4), and ammonia (NH_3). §Comets passing close to the Sun warm considerably, especially their outer layers.

Planets

Contents

Why does the atmospheric content of terrestrial & jovian planets differ?

- **Small planets**: formation via outgassing + planets have low masses & are hot
- **Large planets**: formation via capture/primordial + planets are massive & cold

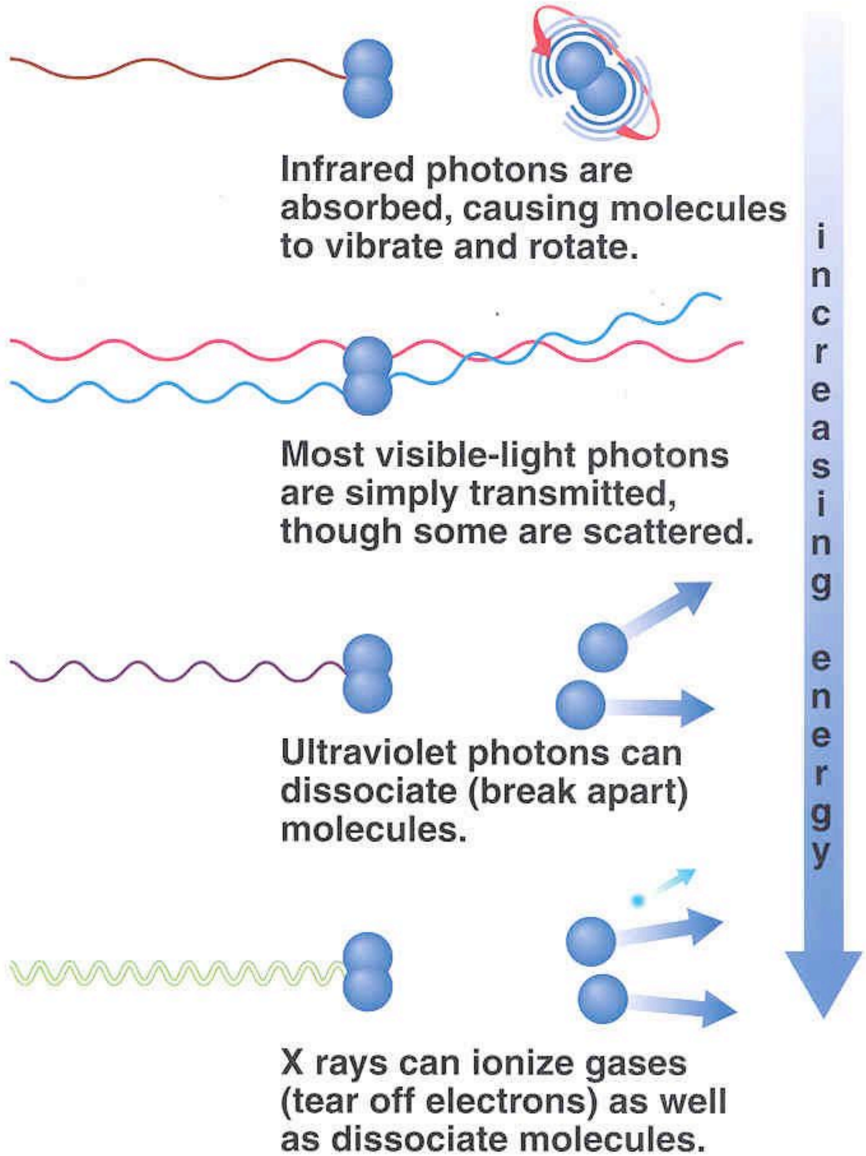
The Generic Atmosphere

- Two Key Points

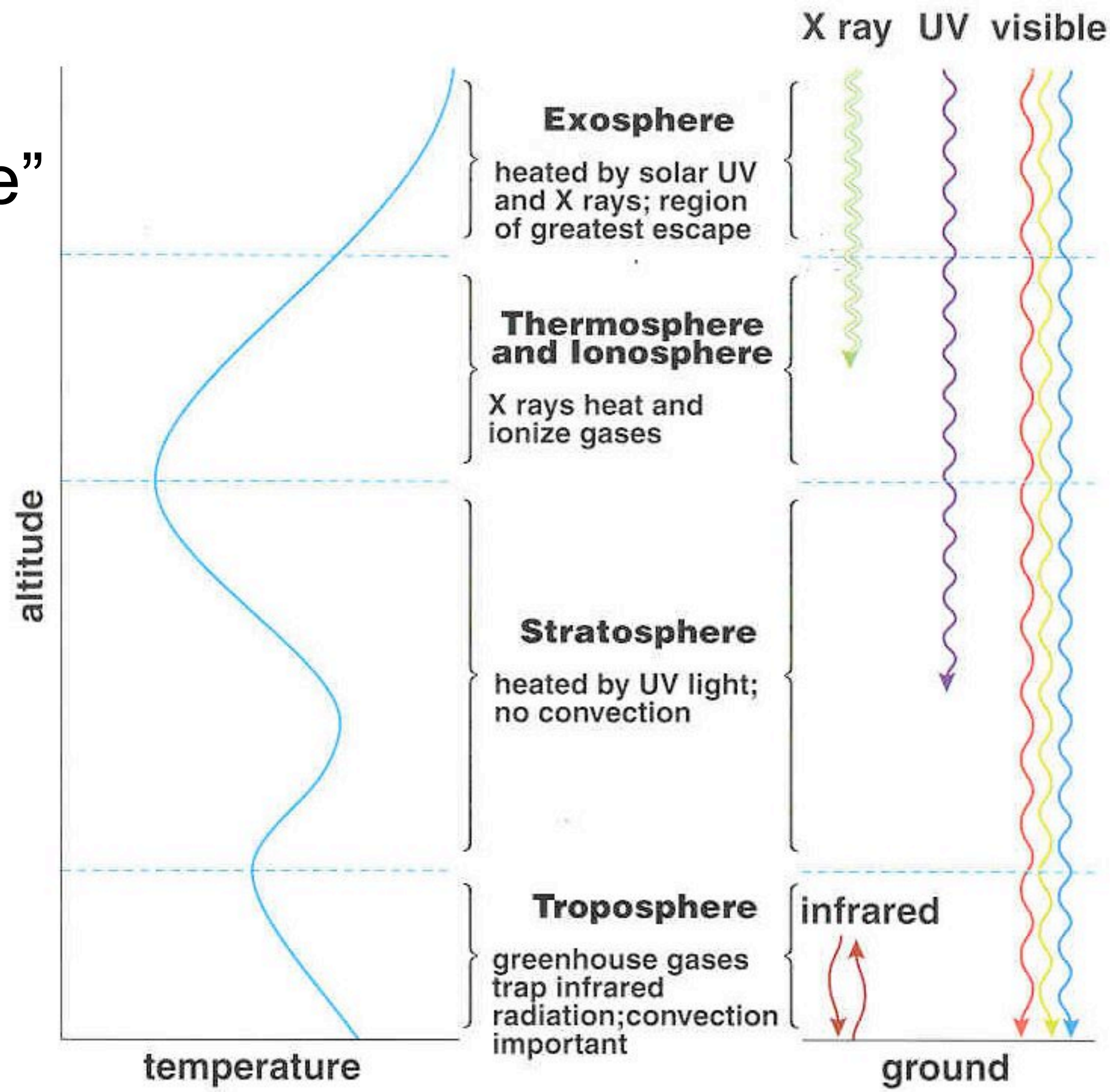
  1) Interaction between the gas & sunlight warms atmosphere
  2) Different wavelengths of light are absorbed in different layers of the atmosphere

- Processes

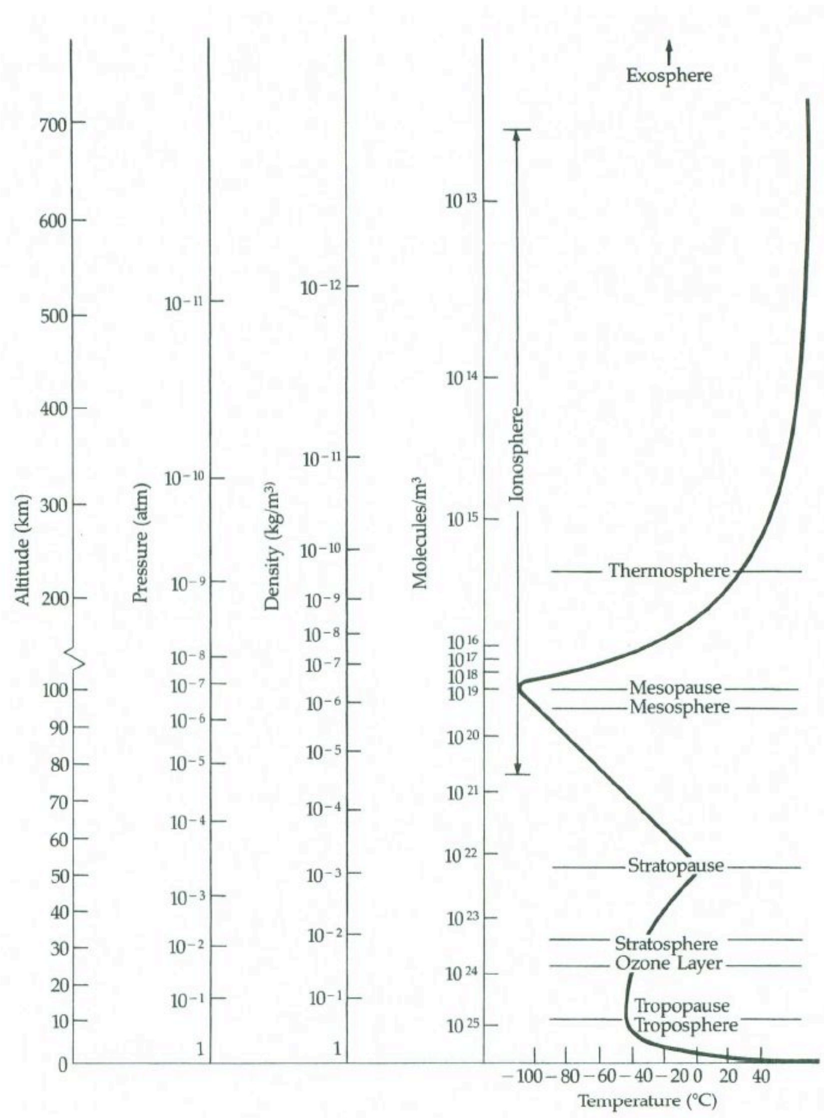
  1) Some gases (greenhouse gases) absorb infrared light very efficiently
  2) Atmosphere is generally transparent to visible light, but can scatter a fraction of visible light
  3) UV photons can break molecules such as Ozone apart
  4) X-rays have enough energy to ionize atoms


Infrared photons are absorbed, causing molecules to vibrate and rotate.
Most visible-light photons are simply transmitted, though some are scattered.
Ultraviolet photons can dissociate (break apart) molecules.
X rays can ionize gases (tear off electrons) as well as dissociate molecules.
increasing energy

“Generic Atmosphere”



Earth: Altitude vs. Temperature

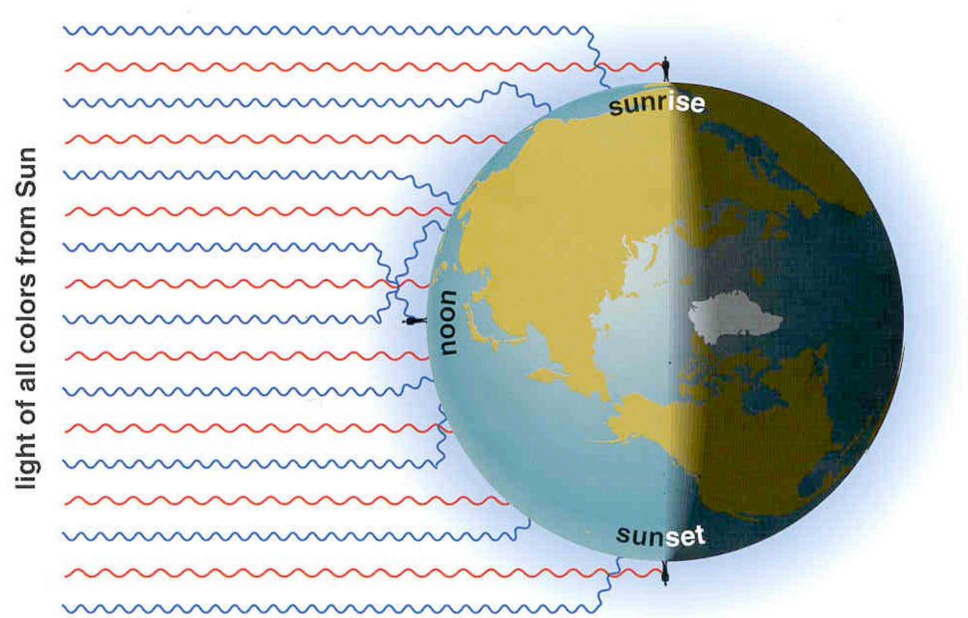


Visible Light

- Absorbed by the Surface
- Shorter wavelength light is easier to scatter than longer wavelength light. The intensity of scattered radiation obeys the **Rayleigh scattering law**,

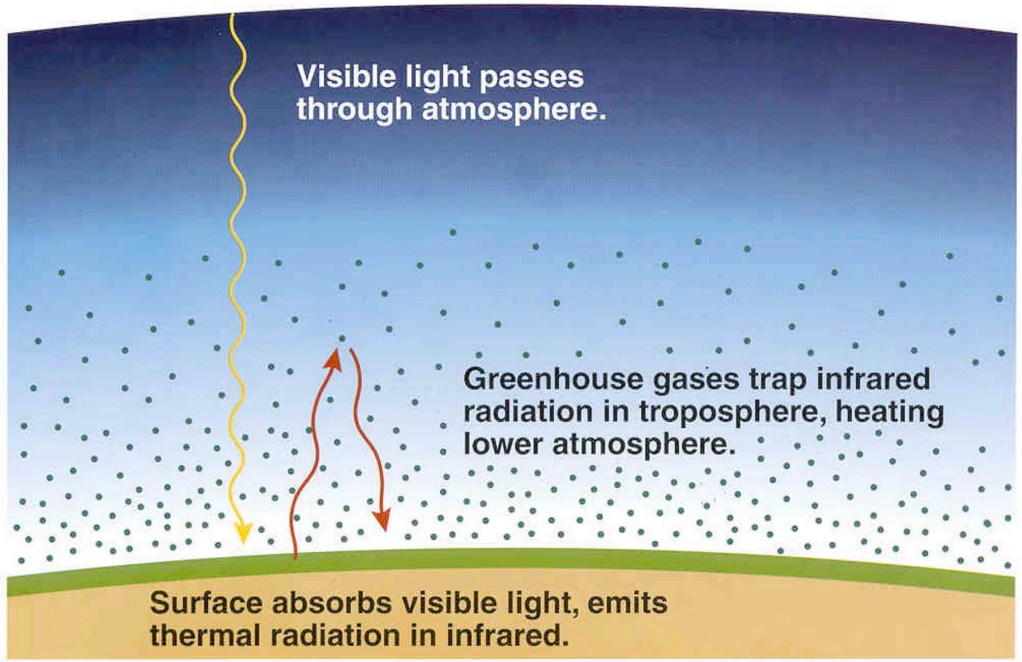
$$I_{\text{scattered}} \propto \frac{1}{\lambda^4}$$

Figure 10.7 Light scattering: blue skies and red sunsets



The Greenhouse Effect

- **Greenhouse Effect**: The process by which greenhouse gases in an atmosphere make a planet's surface temperature warmer than it would be in the absence of an atmosphere
  1) Visible light is absorbed by the surface of a planet
  2) Ground reradiates infrared photons
  3) The infrared photons are absorbed by greenhouse gases (carbon dioxide, water vapor)
  4) Greenhouse gases reradiate infrared photons isotropically
- Where is this important: Troposphere, where the atmosphere is thick
- This warming process causes convection



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*Assumes rapid rotation, in which case day and night temperatures would be the same.

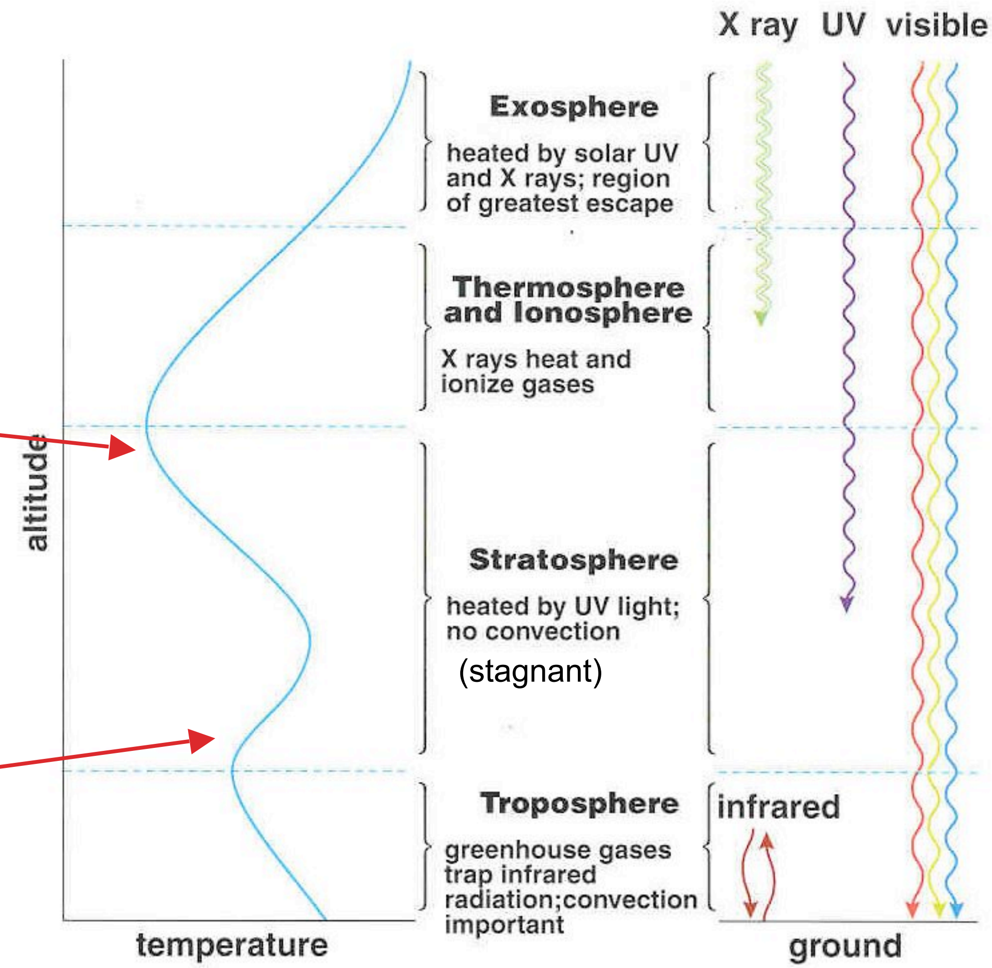
UV & Stratosphere

- Heated by molecular absorption of UV photons. The reaction is

$$O_3 + h\nu_{UV} \rightarrow O_2 + O$$

Heating small because there is no O_3

Heating small because the UV flux is diminished

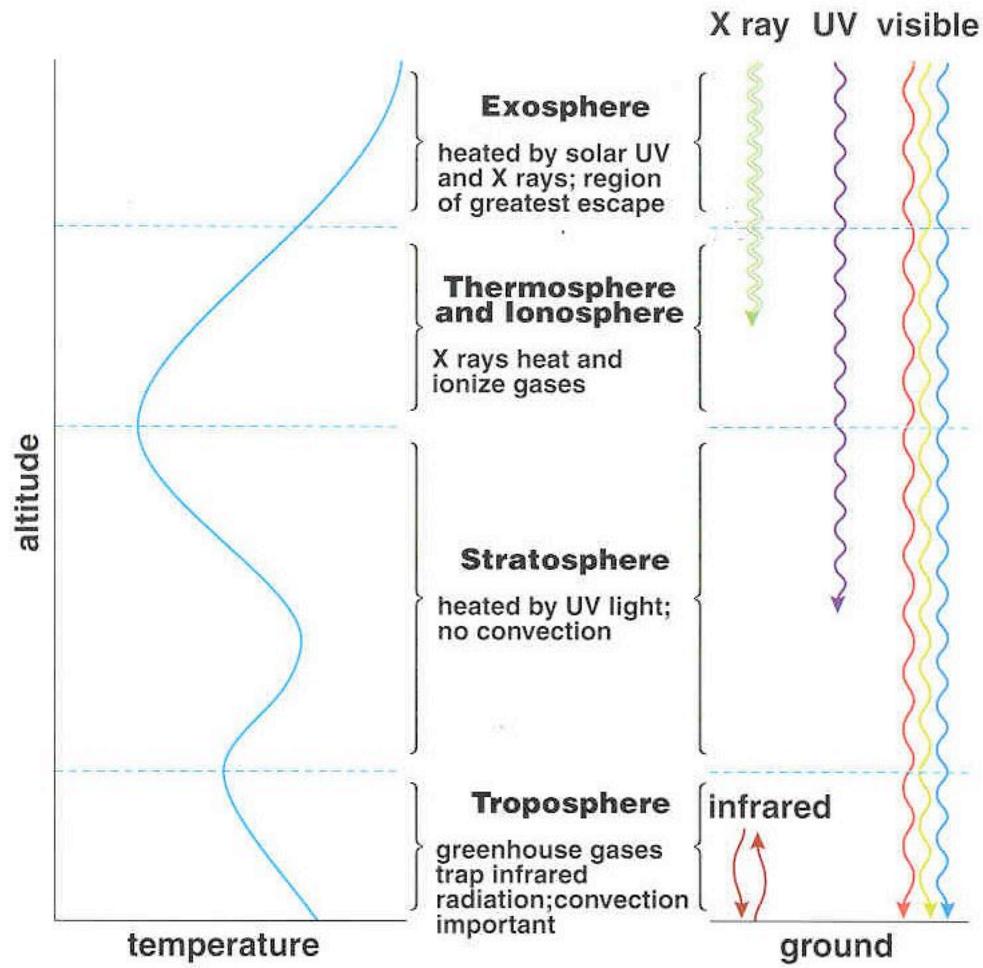


X-rays & Thermosphere

- Origin is similar to the stratosphere, but the heating source is

$$O_2 + h\nu_X \rightarrow O + O$$

- Gas is primarily ions & free electrons
- Because the mass in the thermosphere is small, the resulting temperatures are high (~ 1000-1200 K)



Exosphere

- High temperature, low density gas
- Space Shuttle & artificial satellites orbit the Earth in the exospheric layer
- Drag effects minor but important

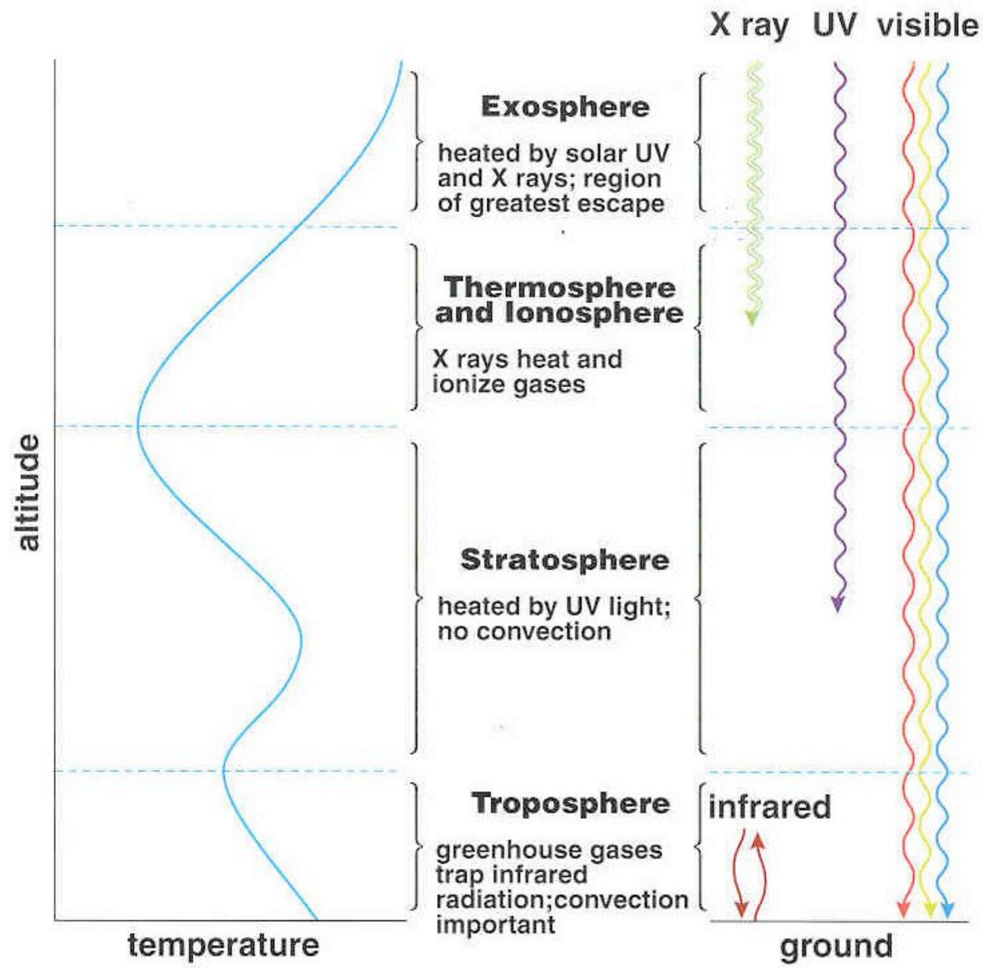
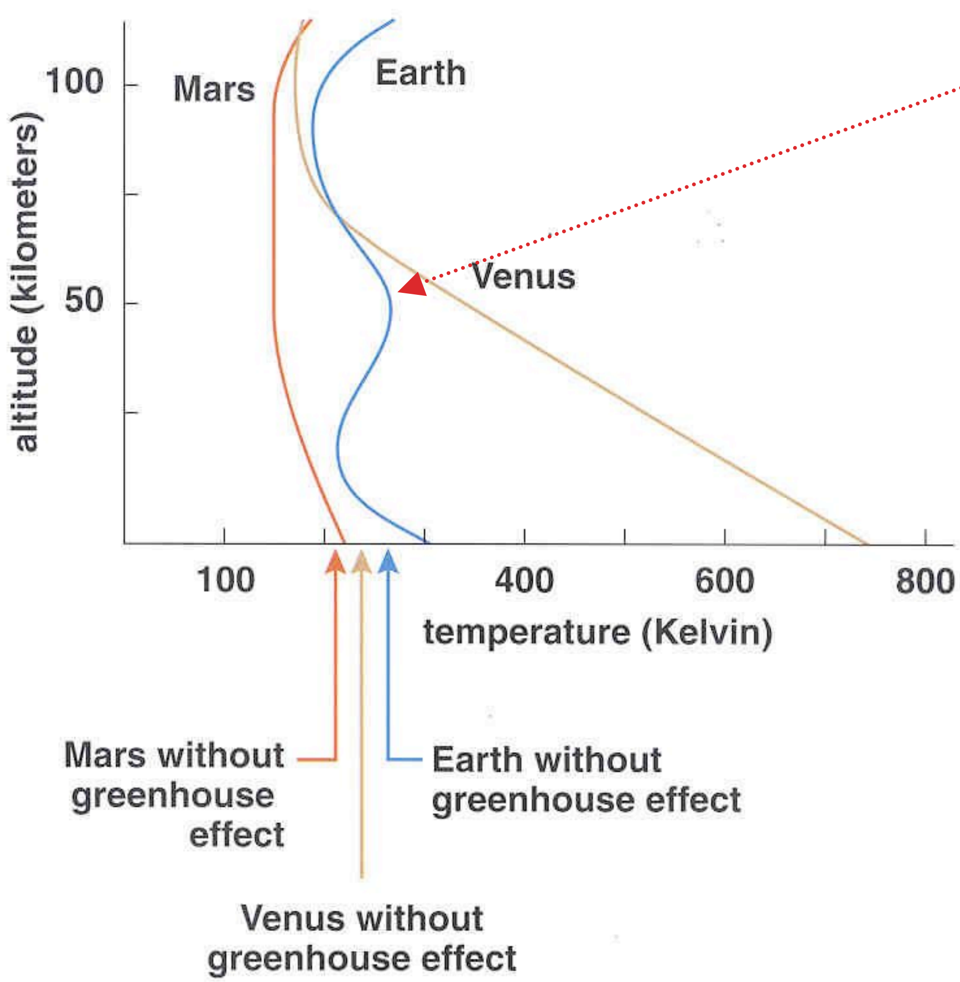
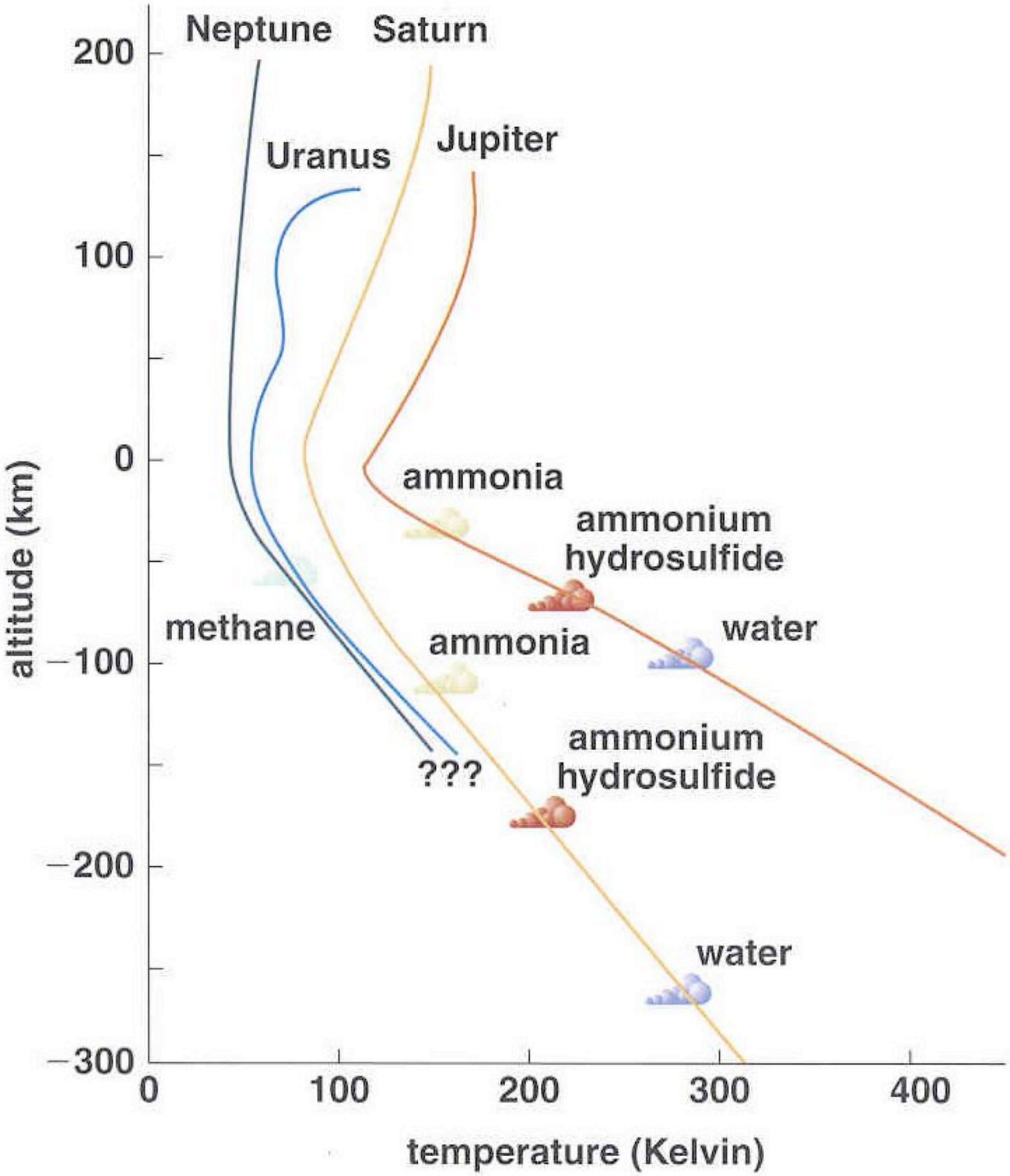


Figure 10.9 Temperature profiles for Venus, Earth, and Mars 76



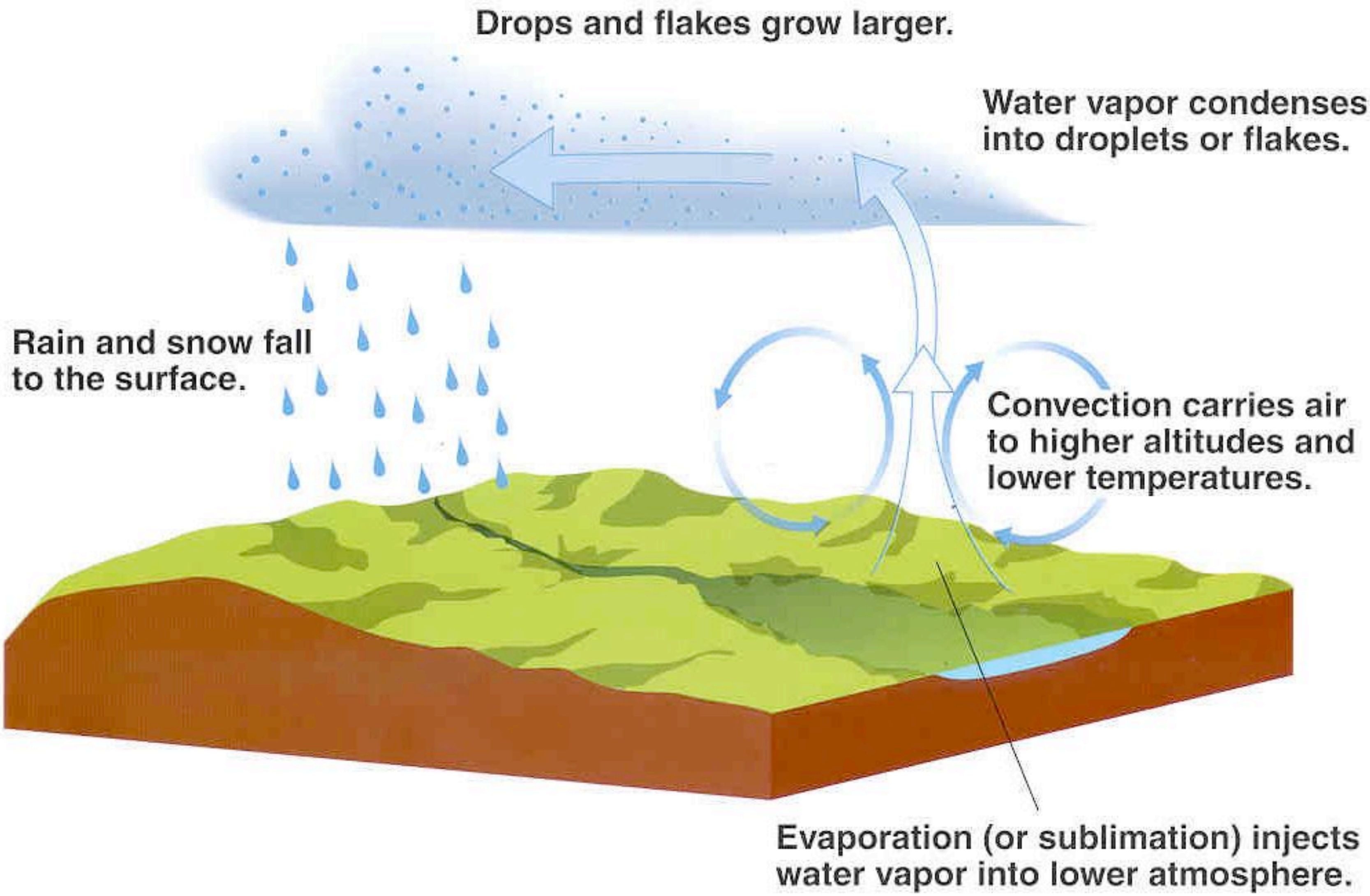
Ozone - note that oxygen readily combines with Si, Al, Mg, etc. to make rocks. The presence of Oxygen our atmosphere indicates that biological processes are producing it.


Neptune
Saturn
Uranus
Jupiter
200
100
0
−100
−200
−300
altitude (km)
ammonia
ammonium
hydrosulfide
water
methane
ammonia
???
ammonium
hydrosulfide
water
0
100
200
300
400
temperature (Kelvin)

Clouds

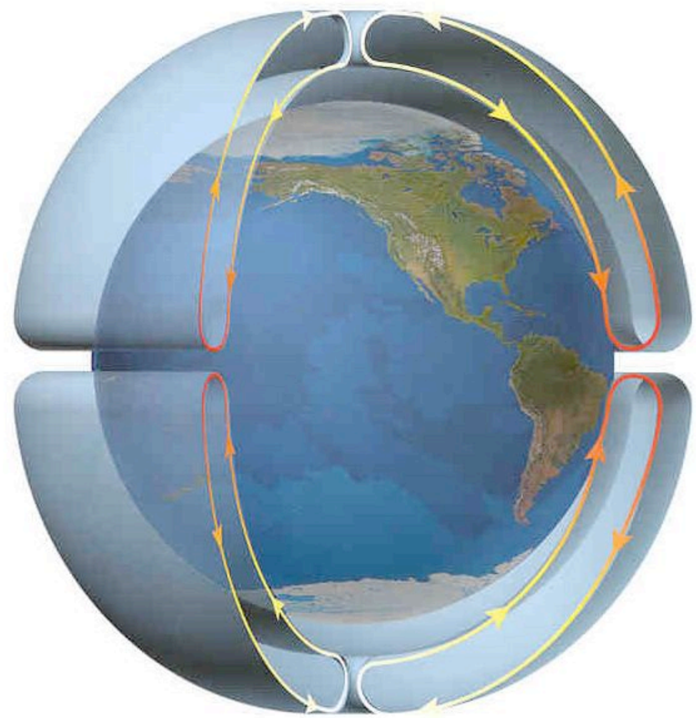
- Forms when one of the gases in the air condenses into liquid/solid form as a result of convection of air ascending a mountainside
- Once droplets grow so large that convection cannot hold them up against gravity, precipitation occurs
- Thus, strong convection → more clouds & precipitation
- Clouds on Different Planets
  → Earth: water
  → Venus: Sulfuric Acid (H_2SO_4) & water droplets

Figure 10.17 Water cycle on Earth



Atmosphere Heating & Wind Patterns

- Equator heated more than poles, creating convective circulation cells



Wind Patterns & Rotation

- Rotation causes air moving on a planet to deviate from a straight line trajectory (**coriolis effect**)

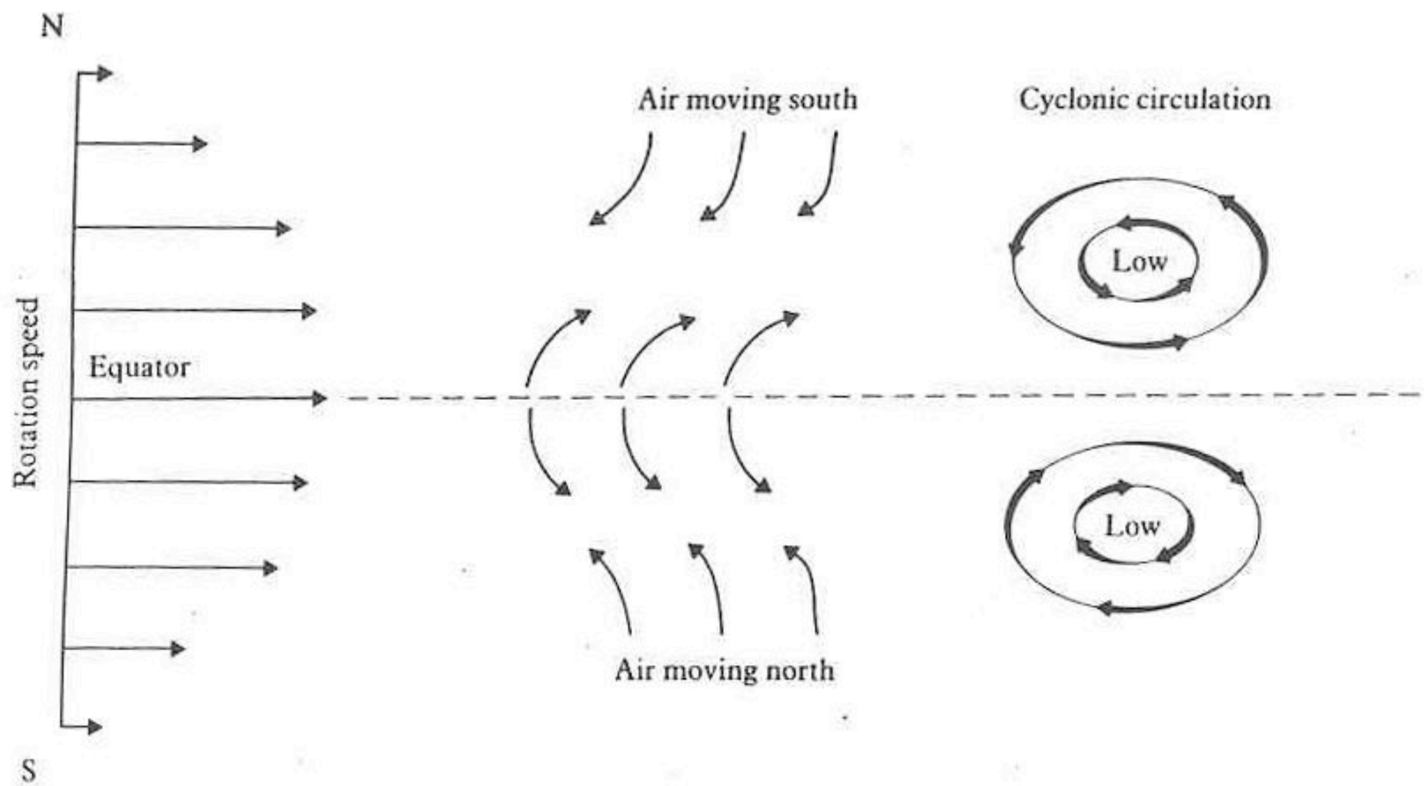


FIGURE 9.17 The Coriolis effect on air moving north or south on a rotating planet. You can experience this same force by trying to walk toward or away from the center of a spinning carousel.

Coriolis Effect

- Rotation causes air moving on a planet to deviate from a straight line trajectory

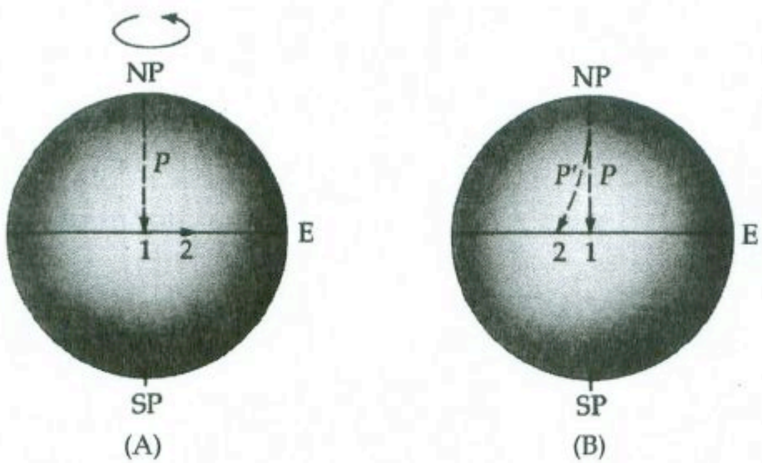
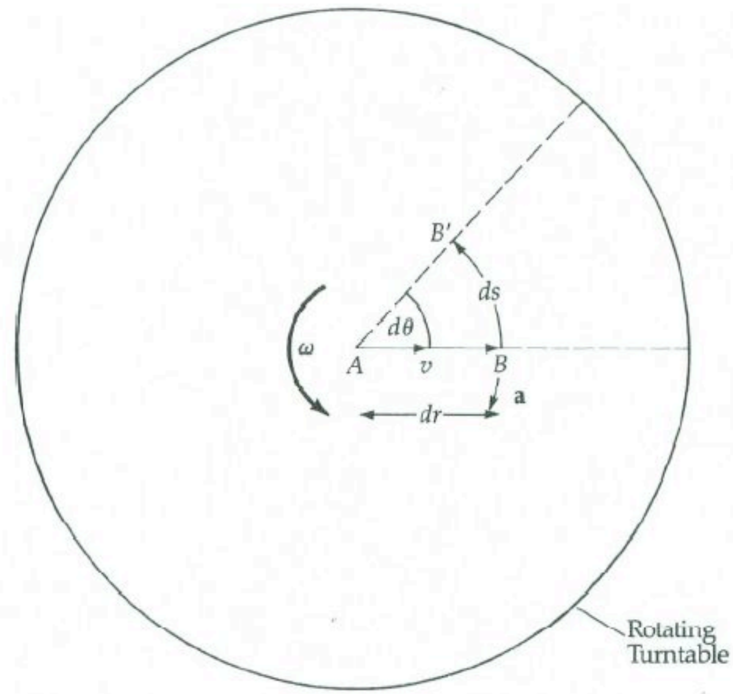


Figure 3–6 Projectile trajectories on the Earth. (A) On a rotating Earth, a rocket launched from the north pole (NP) is aimed at a target located at point 1. During the flight time, the Earth (and the target) rotate from 1 to 2. (B) The view from the Earth's surface shows the rocket's trajectory if the Earth did not rotate (*P*) and the actual one (*P'*). Note that the path curves to the right as seen by an observer at the north pole.

Coriolis Effect

- Consider a turntable rotating with speed ω and a body moving above the table with radial velocity *v*. At time, *t*, the body moves a distance dr from A to B within a time *dt*. Thus,

$$dr = vdt$$



- In *dt*, point B has moved to B'. The angle swept out by this motion is $d\theta$. So,

$$d\theta = \omega dt$$

- Note, also that,

$$d\theta = \frac{ds}{dr} \quad \rightarrow \quad ds = drd\theta$$

Coriolis Effect

- Thus,

$$ds = (dr)(d\theta) = (vdt)(\omega dt) = v\omega(dt)^2$$

- We know that, in general,

$$d = \frac{1}{2}at^2$$

- So, applying this to our turntable,

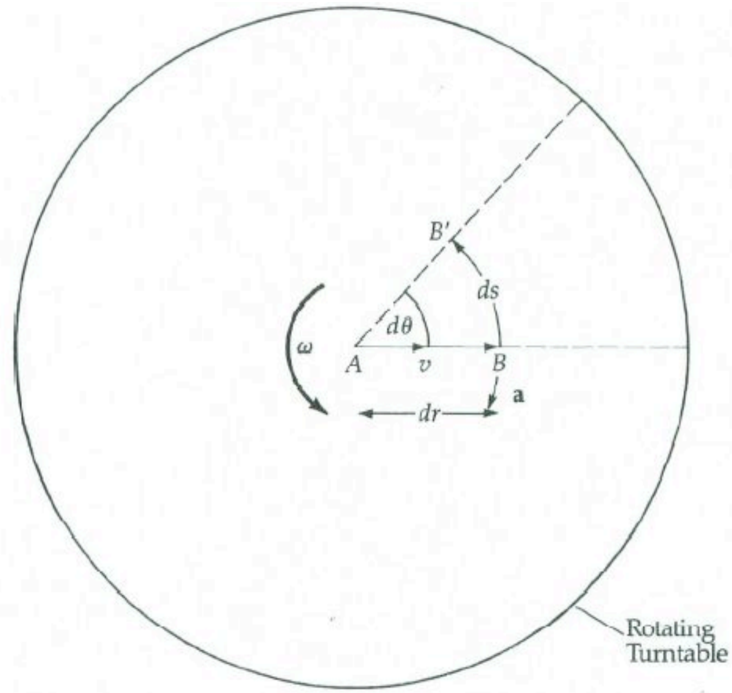
$$ds = \frac{1}{2}a(dt)^2$$

- So,

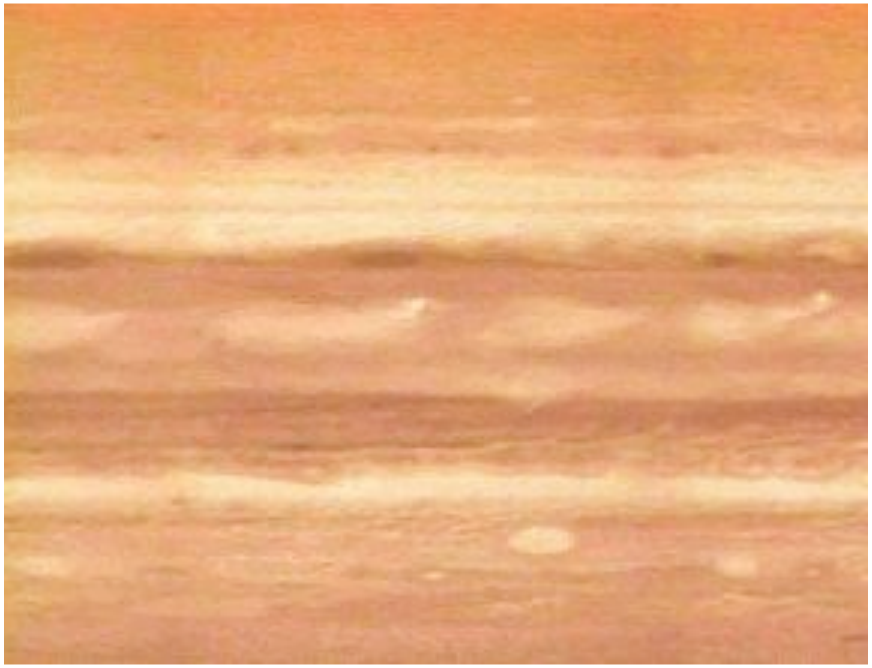
$$a = 2ds\frac{1}{(dt)^2} = 2v\omega(dt)^2\frac{1}{(dt)^2} = 2v\omega$$

- In terms of vectors (the vector ω is towards you),

$$a_{\text{coriolis}} = 2\vec{v} \times \vec{\omega}$$



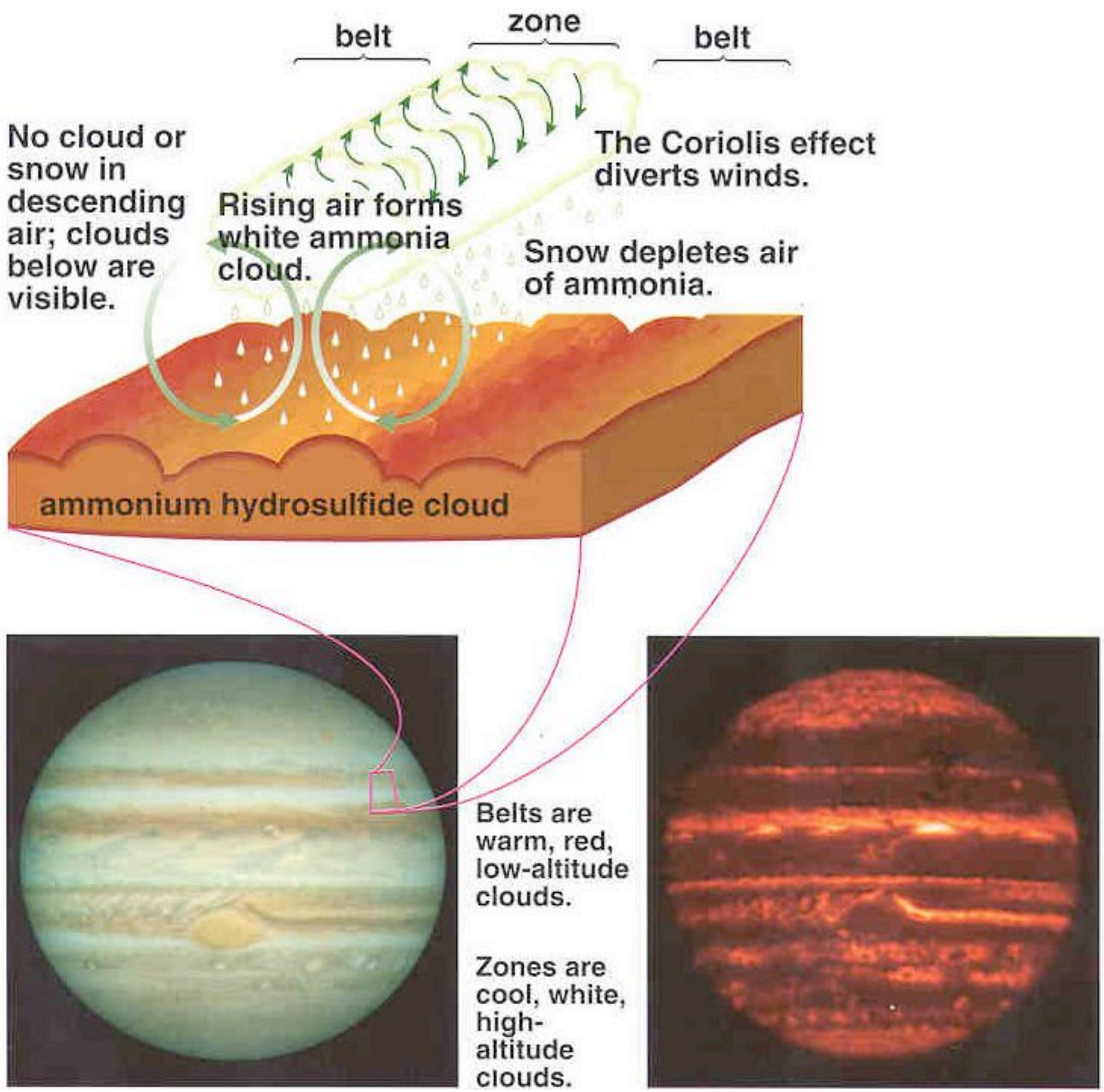
Gas Giants Have Strong Coriolis



- … which divide circulation cells into bands
- Convection results in bands
- For Jupiter -

→ **Zones**: rising, cooling air out of which ammonia condenses into clouds

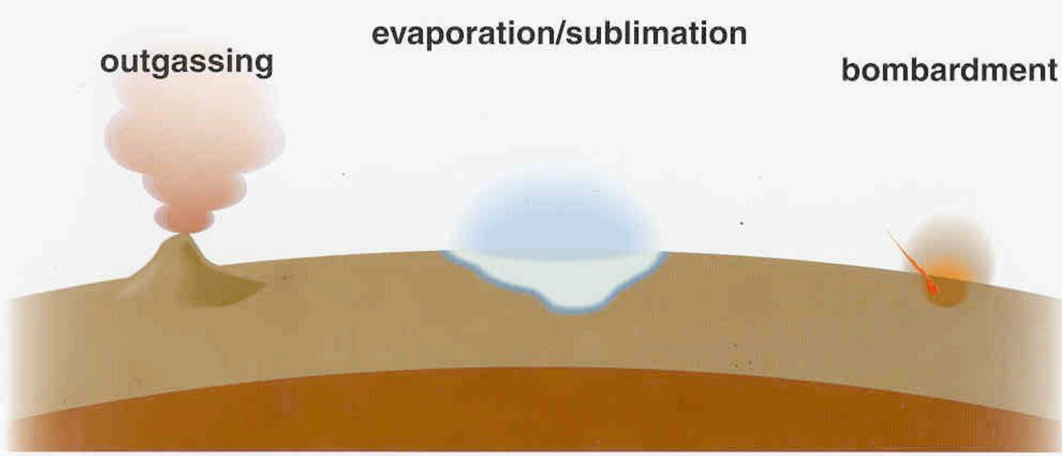
→ **Belts:** falling air depleted in clouds; allows clouds below to be seen


belt
zone
belt
No cloud or snow in descending air; clouds below are visible.
Rising air forms white ammonia cloud.
The Coriolis effect diverts winds.
Snow depletes air of ammonia.
ammonium hydrosulfide cloud
Belts are warm, red, low-altitude clouds.
Zones are cool, white, high-altitude clouds.

Origin & Evolution of Atmosphere

- **Outgassing** – volcanoes expel water, carbon dioxide, molecular nitrogen, hydrogen sulfide, sulfur dioxide
- **Evaporation/Sublimation** – important for Mars
- **Bombardment** – important for Mercury & the Moon

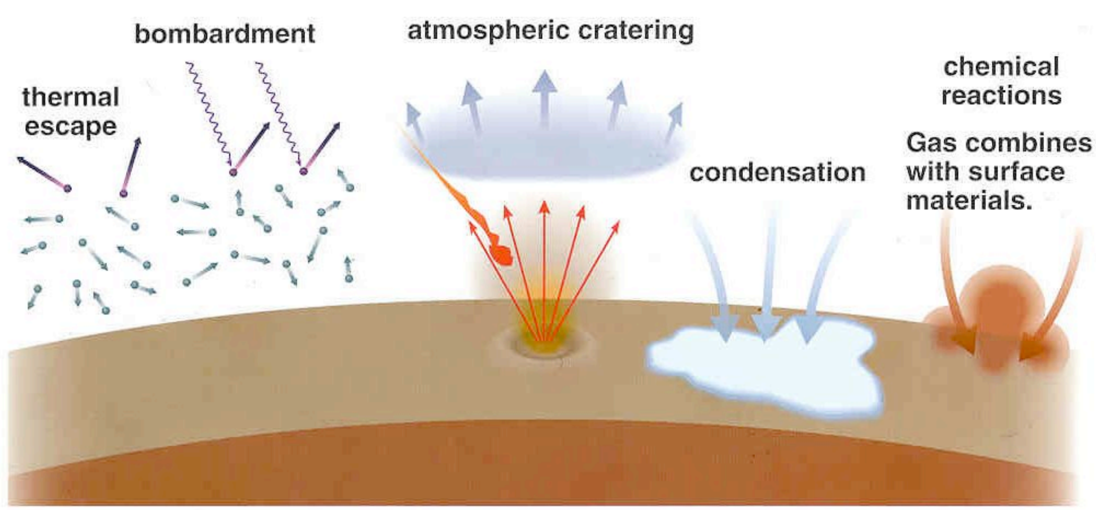
Figure 10.20 Three processes by which atmospheres gain gas 82



How are gases lost?

1) **Thermal Escape** – atoms/ molecules moving fast enough to escape the pull of gravity
2) **Bombardment**: gives kinetic energy (speed) to atmospheric atoms & molecules & breaks heavier molecules into lighter atoms

Figure 10.22 The five major processes by which atmospheres gain gas   83



Losing gases, cont.

3) **Atmospheric cratering** – via large impacts is important on small worlds

4) **Condensation** – via cooling is important for Mars & polar craters of the Moon and Mercury

5) **Chemical Reactions**: air combining with rocks

Figure 10.22 The five major processes by which atmospheres gain gas