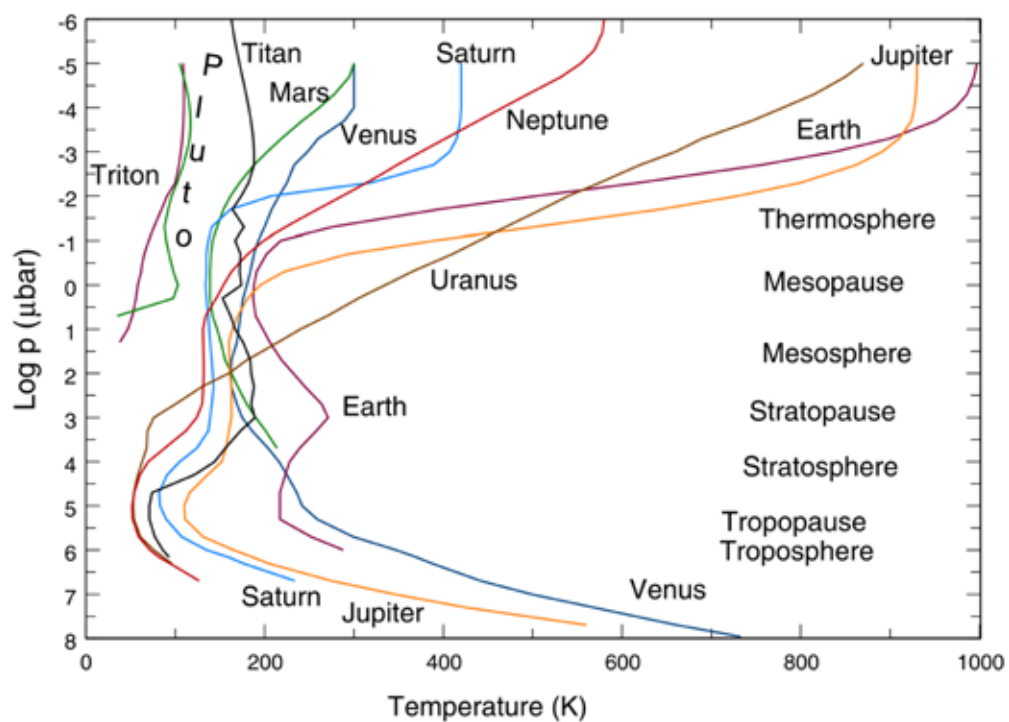


# Vertical structure of the atmosphere

Thermodynamics, Radiative transfer, Radiative-convective equilibrium

## Vertical structures of planetary atmospheres



(Mueller-Wodarg et al.)

# Hydrostatic equilibrium

The gravitational acceleration is assumed to be a constant value  $g$ .

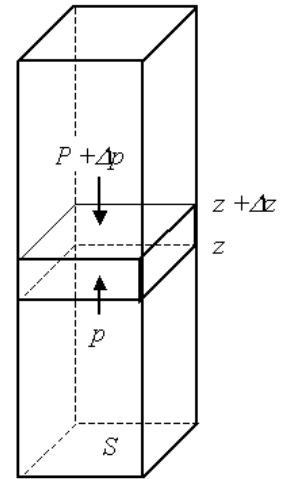
The balance between the pressure gradient force and the gravitational acceleration in the vertical direction is

$$\begin{aligned} -S\Delta p &= g\rho S\Delta z \\ \therefore \frac{dp}{dz} &= -g\rho \end{aligned} \quad (1.1)$$

$p$ : pressure    $z$ : altitude    $\rho$ : mass density (kg/m<sup>3</sup>)

This is equivalent to the vertical momentum equation

$$\frac{dw}{dt} = -\frac{1}{\rho} \frac{\partial p}{\partial z} - g$$



except that the vertical wind  $w$  is assumed to be zero.

Integrating (1.1) we have

$$p(z) = \int_z^\infty g\rho(z')dz'$$

The equation of state:

$$p = \rho RT \quad (1.2)$$

$R$ : gas constant (= 287 J/K/kg for Earth)

$R = k/m$ , where  $k$  is Boltzmann's constant and  $m$  is the mean mass of molecules

Combining (1.1) (1.2), we have

$$\frac{dp}{dz} = -\frac{gp}{RT}$$

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

$p_s$ : surface pressure

$H = RT/g$ : scale height (6-8km on Earth)  
(~16 km on Venus, ~11 km on Mars)

When the temperature is constant with altitude,

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

Atmospheres at different altitudes do not readily mix.  
Horizontally-thin layers tend to prevail.

← Air parcels tend to stay at constant levels in a stably-stratified atmosphere



Cumulonimbus clouds



Haze layer of Titan

## Thermodynamics

First law of thermodynamics:

$$dH = c_v dT + p d\alpha \quad (2.1)$$

$dH$  : heat given to gas of unit mass

$c_v$  : specific heat for constant volume

$\alpha = 1/\rho$  : specific volume

Combined with the state equation  $p\alpha = RT$ , we obtain

$$p d\alpha + \alpha dp = R dT \quad (2.2)$$

Combining (2.1)(2.2) yields

$$dH = c_p dT - \alpha dp \quad (2.3)$$

$c_p = c_v + R$  : specific heat for  
constant pressure

Considering an adiabatic process ( $dH = 0$ ), we have

$$c_p dT = \frac{RT}{p} dp$$

$$c_p d(\ln T) = R d(\ln p)$$

$$\therefore T = \text{const.} \times p^{R/C_p}$$

Then, the potential temperature (温位)  $\theta$  defined as

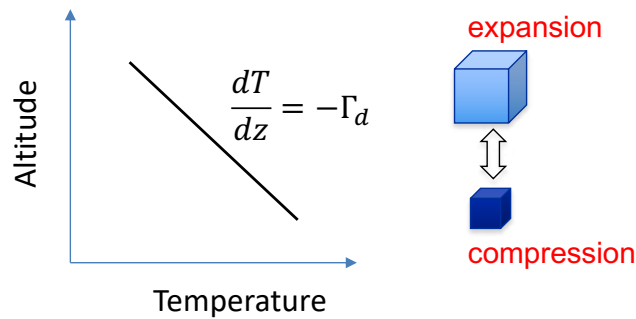
$$\theta \equiv T \left( \frac{p_s}{p} \right)^{R/C_p} \quad (2.4)$$

is conserved in adiabatic processes.

In adiabatic ascent or decent,  $\theta$  is constant with altitude. In this case, under hydrostatic equilibrium, we obtain

$$\frac{dT}{dz} = -\frac{g}{c_p} = -\Gamma_d \quad (2.5)$$

$\Gamma_d = g/C_p$  : dry adiabatic lapse rate (9.8 K/km on Earth)



**Table 1.3** Thermodynamic properties and lapse rates on various planets.

Body	Main gases	Molar mass, $M$ g mol <sup>-1</sup>	$\bar{R}$ J kg <sup>-1</sup> K <sup>-1</sup>	$g$ m s <sup>-2</sup>	$c_p$ J kg <sup>-1</sup> K <sup>-1</sup>	$\Gamma_a = g/c_p$ troposphere K km <sup>-1</sup>	$\Gamma$ observed in troposphere K km <sup>-1</sup>
Venus	CO <sub>2</sub> , N <sub>2</sub>	43.45	189	8.901	930	9.5	~8.0
Earth	N <sub>2</sub> , O <sub>2</sub>	28.97	287	9.81	1004	9.8	~6.5
Mars	CO <sub>2</sub> , N <sub>2</sub>	43.5	191	3.72	850	4.4	~2.5
Jupiter	H <sub>2</sub> , He	2.22	3745	24.25	10 988	2.1–2.45*	1.9
Saturn	H <sub>2</sub> , He	2.14	3892	10.0	10 658	0.7–1.1	0.85
Titan	N <sub>2</sub> , CH <sub>4</sub>	28.67	290	1.36	1044	1.3	1.0–1.4
Uranus	H <sub>2</sub> , He	2.3	3615	8.80	8643	0.7–1.1	0.75
Neptune	H <sub>2</sub> , He	2.3	3615	11.1	8643	0.85–1.34	0.95
HD209458b	H <sub>2</sub>	2.0	4160	18.5	14 300	1.3	~0.2 (model)**

\* On giant planets,  $g$  varies from equator to pole.

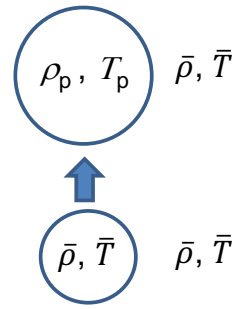
\*\* Menou and Rauscher (2009).

# Static stability

The buoyancy acting on an air parcel is given by

$$\frac{d^2 z}{dt^2} = g \frac{\bar{\rho} - \rho_p}{\rho_p}$$

$z$  : altitude of the air parcel  
 $\bar{\rho}$  : mass density of ambient air  
 $\rho_p$  : mass density of the air parcel



Assuming that the pressures of the air parcel and the ambient atmosphere are equal, we have

$$\frac{d^2 z}{dt^2} = g \frac{\bar{T}^{-1} - T_p^{-1}}{T_p^{-1}} = g \frac{T_p - \bar{T}}{\bar{T}} \quad (2.6)$$

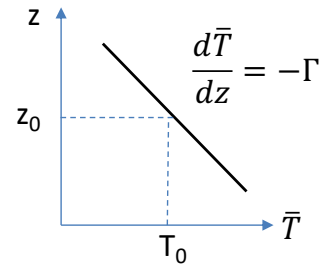
$\bar{T}$  : ambient temperature  
 $T_p$  : temperature of air parcel

Temperatures are expressed by using the temperature  $T_0$  at the original position ( $z = 0$ ):

$$\begin{aligned} \bar{T} &= T_0 - \Gamma z \\ T_p &= T_0 - \Gamma_d z \end{aligned} \quad (2.7)$$

$\Gamma = -d\bar{T}/dz$  : ambient lapse rate

$\Gamma_d = g/C_p$  : adiabatic lapse rate



From (2.6)(2.7)

$$\frac{d^2 z}{dt^2} \sim -g \frac{\Gamma_d - \Gamma}{T_0} z$$

When  $\Gamma_d - \Gamma$  is positive, an oscillating solution exists.

The buoyancy frequency  $N$  is given by

$$N^2 = g \frac{\Gamma_d - \Gamma}{T_0} = g \frac{\partial \ln \bar{\theta}}{\partial z}$$

Static stability is defined by

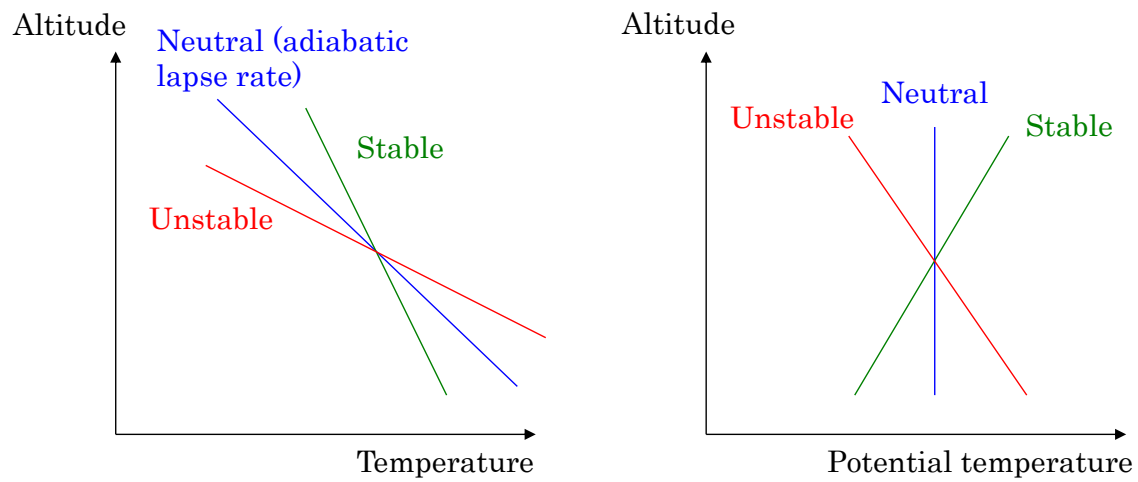
$$S = \frac{dT}{dz} + \Gamma_d$$

Three types of static stability condition:

$$\begin{aligned} \Gamma_d - \Gamma > 0 &\leftrightarrow S > 0 \leftrightarrow \partial \bar{\theta} / \partial z > 0 && : \text{stable} \\ \Gamma_d - \Gamma = 0 &\leftrightarrow S = 0 \leftrightarrow \partial \bar{\theta} / \partial z = 0 && : \text{neutral} \\ \Gamma_d - \Gamma < 0 &\leftrightarrow S < 0 \leftrightarrow \partial \bar{\theta} / \partial z < 0 && : \text{unstable} \end{aligned}$$

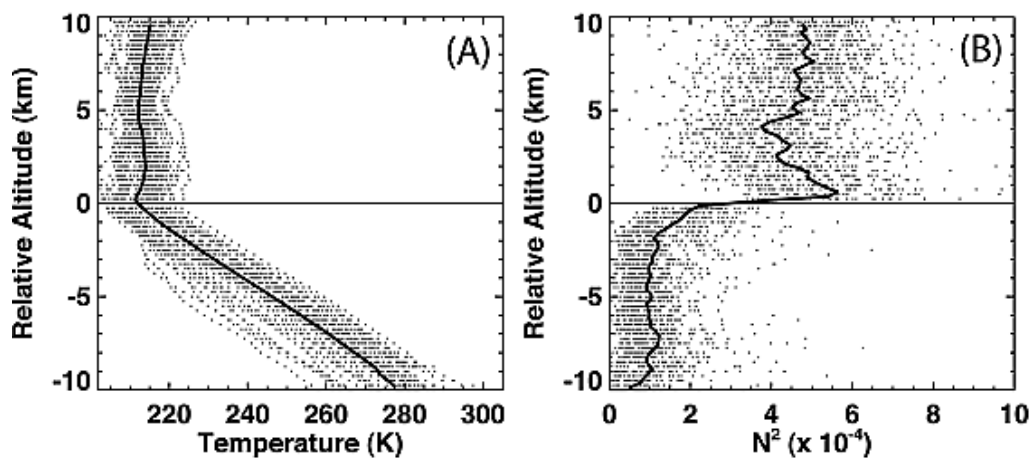
$\Gamma = -d\bar{T}/dz$  : ambient lapse rate

$\Gamma_d = g/C_p$  : adiabatic lapse rate



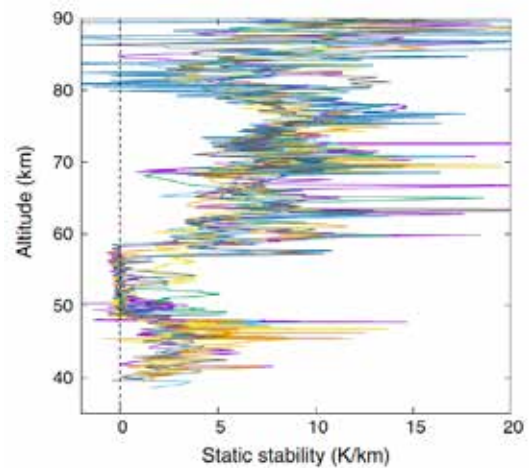
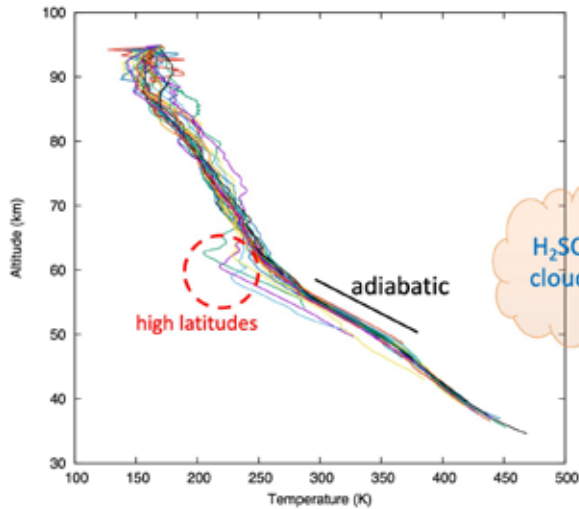
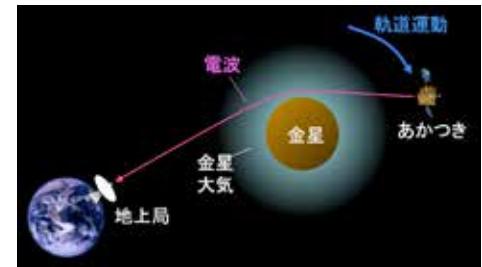
## Stability of Earth's atmosphere

Gettelman et al. (2011)



## Stability of Venusian atmosphere

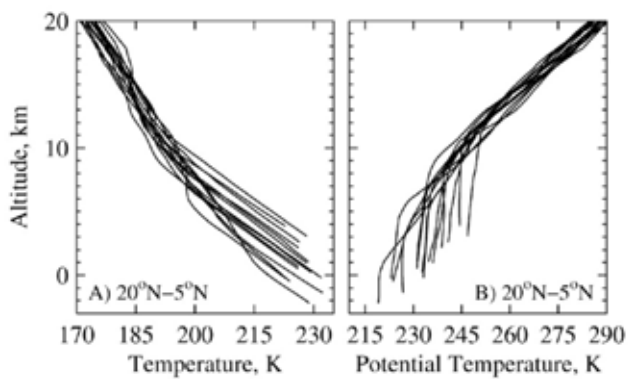
Neutral layer exists around 50-60 km in the cloud



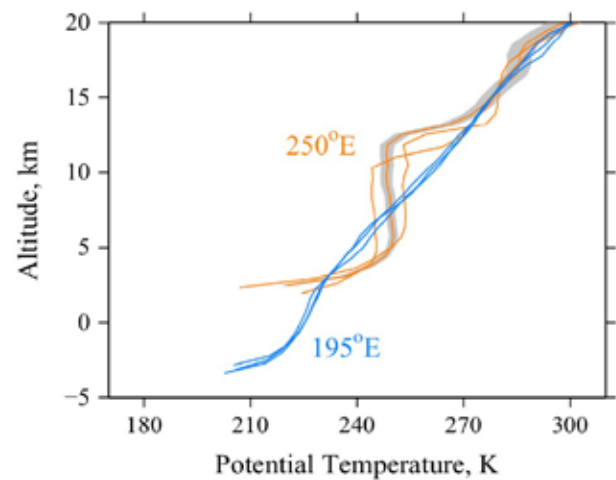
Akatsuki radio occultation (Imamura et al. 2017)

## Stability of Martian atmosphere

Mixed boundary layer (Hinson et al. 2008)



Detached mixed layer (Hinson et al. 2014)

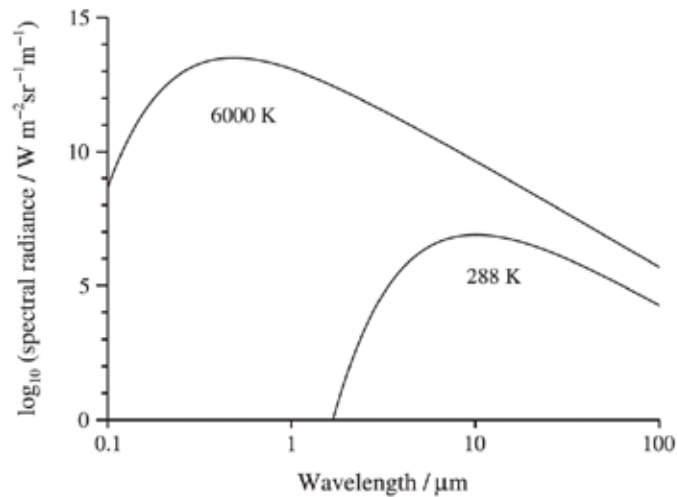


# Radiation

Energy balance of a planet

Inflow : Visible wavelength radiation from Sun

Outflow : Infrared radiation from the surface and the atmosphere



Andrews (2000)

Logarithm of the black-body spectral radiance  $B_\lambda(T)$ , plotted against the logarithm of wavelength  $\lambda$ , for  $T = 6000$  K, a typical temperature of the solar photosphere, and 288 K, the Earth's mean surface temperature.

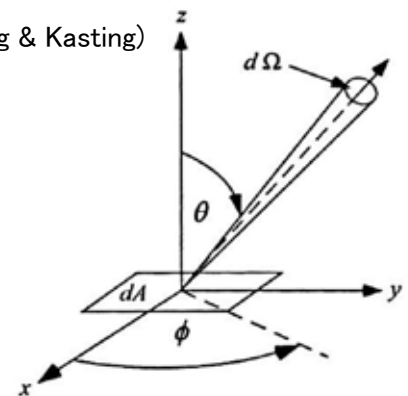
Planck function (J/m<sup>2</sup>/s/str/Hz)

$$B_\nu(T) = \frac{2h\nu^3}{c^2(e^{h\nu/kT} - 1)}$$

h: Planck's constant  
v: frequency

c: speed of light  
k: Boltzmann's constant

(Catling & Kasting)



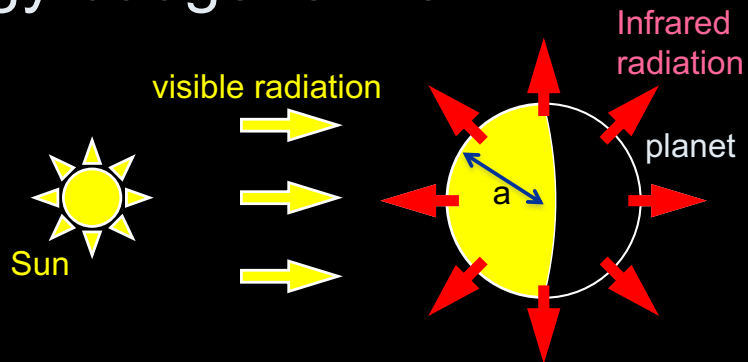
Integration for wavelength and for solid angle over a hemisphere

$$\underbrace{\int_0^{2\pi} d\phi \int_0^{\pi/2} d\theta \sin\theta \cos\theta}_{\text{Integration for solid angle}} \underbrace{\int_0^\infty d\nu B_\nu(T)}_{\text{Integration for frequency}} = \pi \int_0^\infty d\nu B_\nu(T) = \sigma T^4$$

$\sigma$ : Stefan-Boltzmann's constant



# Energy budget of Earth



Incident solar flux

$$(1 - A) S \pi a^2$$

Outgoing infrared flux

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

$A$ : Albedo (0.3 for Earth)

$S$ : Solar constant ( $1370 \text{ W m}^{-2}$  for Earth)

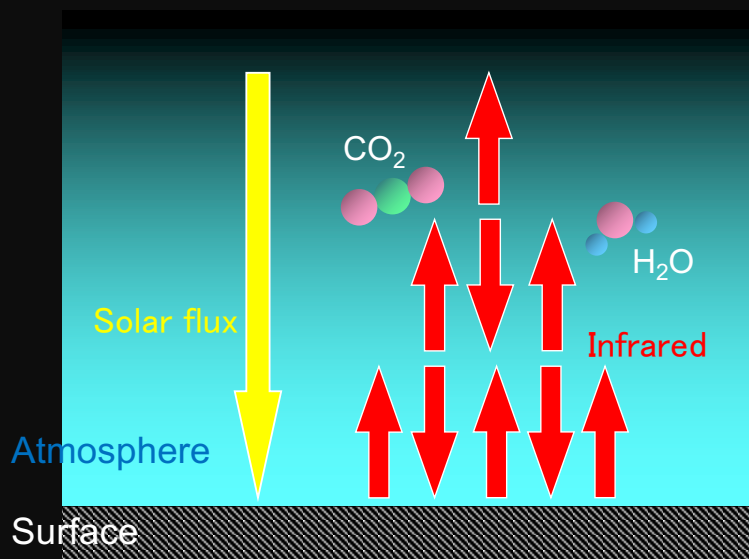
$a$ : Planetary radius

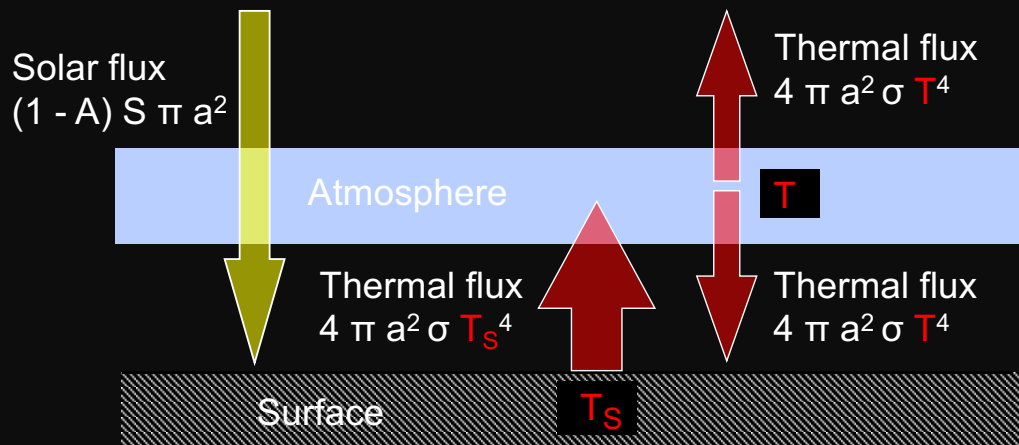
$\sigma$ : Stefan-Boltzmann constant

$T$ : Surface temperature

Substituting the values for Earth, we obtain  $T = 255 \text{ K}$

- In reality the mean surface temperature is around  $15^\circ\text{C}$   
→ Greenhouse effect by water vapor and  $\text{CO}_2$
- Infrared-active gases absorb thermal infrared radiation emitted from the surface and prevents escape of energy to space





Energy balance

Top of atmosphere:  $(1 - A) S \pi a^2 = 4 \pi a^2 \sigma T^4$

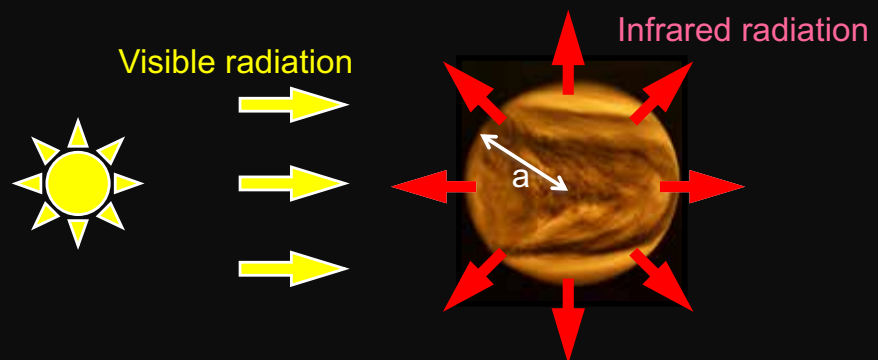
Surface:  $(1 - A) S \pi a^2 + 4 \pi a^2 \sigma T^4 = 4 \pi a^2 \sigma T_s^4$

Solving these, we obtain

$$T_s = \sqrt[4]{2} \times T = 1.19 T$$

$T$  is equal to the surface temperature in the absence of the atmosphere. The surface temperature  $T_s$  increases as a result of the “greenhouse” effect.

## Energy budget of Venus



$$(1 - A) S \pi a^2 = 4 \pi a^2 \sigma T^4$$

A: Albedo (0.78)

S: Solar flux (2617 W m<sup>-2</sup>)

Substituting the values for Venus, we obtain  $T = 224 \text{ K}$

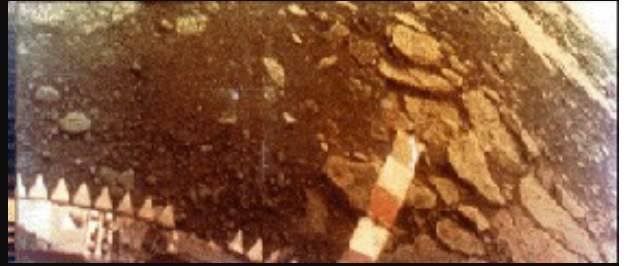
In spite of the proximity to the Sun, the energy input to Venus is low due to the reflective cloud. Venus should freeze.

## Greenhouse effect on Venus ?

- Atmospheric composition

CO<sub>2</sub> 96.5 %

N<sub>2</sub> 3.5 %



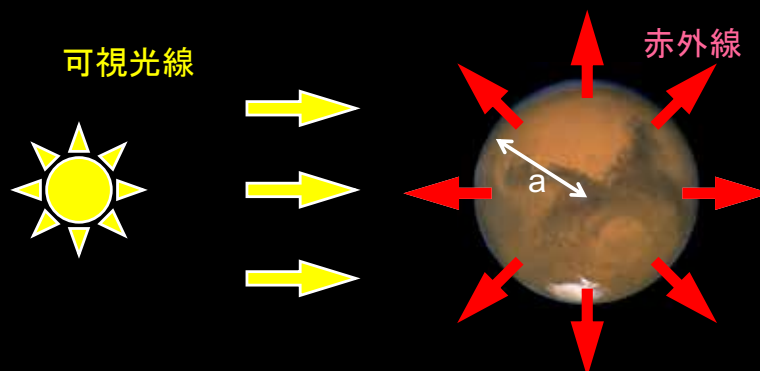
- Surface pressure = 92 atm

Thick atmosphere

The amount of CO<sub>2</sub> is 300,000 times larger than that of Earth

Greenhouse effect increases the surface temperature up to 460° C

## Energy budget of Mars



$$(1 - A) S \pi a^2 = 4 \pi a^2 \sigma T^4$$

A: Albedo (0.16)

S: Solar flux (589 W m<sup>-2</sup>)

Substituting the values for Mars, we obtain  $T = 216 \text{ K}$

In spite of the low albedo, the large distance from the Sun leads to a freezing world

# Greenhouse effect on Mars ?

- Atmospheric composition

CO<sub>2</sub> 95 %

N<sub>2</sub> 2.7 %

Ar 1.6 %

- Pressure 0.006 atm

Thin atmosphere

Water vapor, which also plays a major role in the greenhouse effect on Earth, is also scarce.



**Greenhouse effect on Mars is as small as several Kelvins.**

## Energy budget and temperature

	Solar flux (W/m <sup>2</sup> )	Albedo	Absorbed energy (W/m <sup>2</sup> )	Greenhouse effect	
				off	on
Venus	2617	0.78	576	-50°C	470 °C
Earth	1370	0.30	959	-18°C	15 °C
Mars	589	0.16	495	-57°C	-53°C

# Interaction between electromagnetic waves and molecules

スピンの変化		姿勢の変化	原子配置 の変化	電子分布の変化		核配置 の変化
核磁気共鳴	電子スピン 共鳴	マイクロ波	赤外	紫外・可視	X-線	$\gamma$ -線
$10^{-2}$	1	100	$10^4$	$10^6$	波数	$10^8$
10 m	100 cm	1 cm	100 $\mu$ m	1 $\mu$ m	10 nm	波長
$3 \times 10^8$	$3 \times 10^8$	$3 \times 10^{10}$	$3 \times 10^{12}$	$3 \times 10^{14}$	Hz	周波数
$10^{-3}$	$10^{-1}$	10	$10^3$	$10^5$	joules/mole	エネルギー

図 5.1 電磁波のスペクトルと電磁波-分子 (原子) の相互作用のメカニズム  
(Banwell and McCash, 1994)<sup>26)</sup>

柴田 (1999)

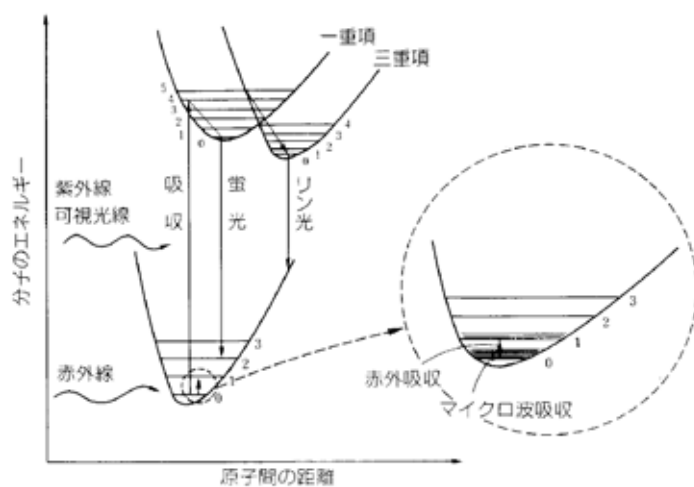
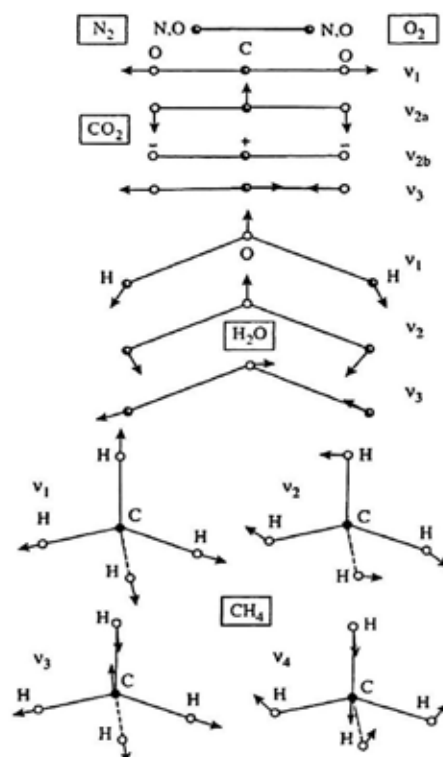


図 1-3 分子のエネルギーレベルと光の吸収

柴田 (1999)



Catling & Kasting (2017)

Vibrational energy levels

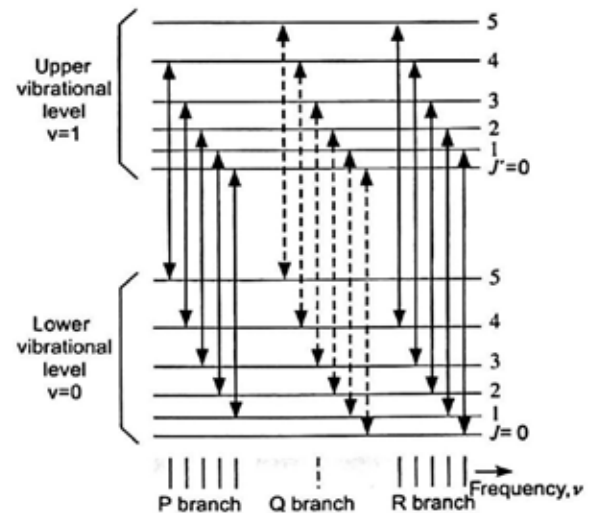
$$E_v = h\nu_0 \left( v + \frac{1}{2} \right)$$

$v$  : vibrational quantum number

Rotational energy levels

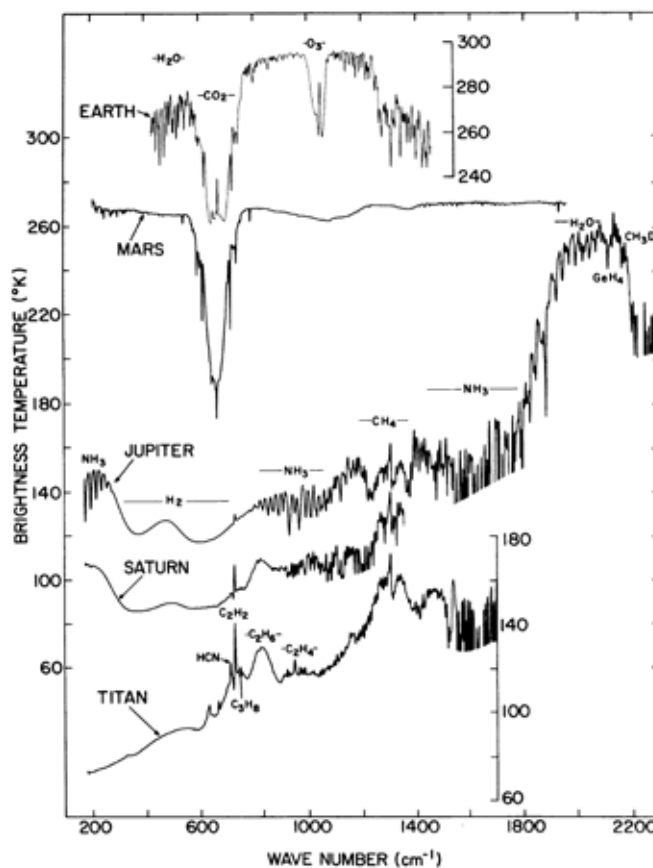
$$E_J = hB(J(J + 1))$$

$J$  : rotational quantum number



**Figure 2.23** The meaning of the P, Q, and R branches in vibrational-rotational transitions, showing a vibrational transition  $\Delta v$  and superposed rotational transitions,  $\Delta J$ . The P branch corresponds to  $\Delta v = 1$  with rotational transitions  $\Delta J = -1$ . The Q branch corresponds to  $\Delta v = 1$  and  $\Delta J = 0$  and the R branch corresponds to  $\Delta v = 1$  and  $\Delta J = +1$ . The lower shaded panel shows the appearance of the lines in the spectrum schematically, noting that the Q branch is offset from P and R branches for clarity in order to show the Q-branch  $\Delta J$  transitions.

Catling & Kasting (2017)



**Fig. 4.21** Spectra in the thermal infrared, plotted as brightness temperatures, for four planets and Titan. Features that show as "absorptions" are formed in a region of negative temperature gradient (troposphere); those that show as "emissions" are from a warm stratosphere. [After HANEL (1983).]

Chamberlain & Hunten (1987)

# Radiative transfer in plane-parallel atmosphere

Radiative transfer equation

$$dI = (\text{absorption}) + (\text{emission}) = -k_a I ds + k_a B(T) ds$$

$$\therefore \frac{dI}{k_a ds} = -I + B(T)$$

$I$ : radiance (J/m<sup>2</sup>/s/str/Hz)

$k_a$ : absorption coefficient

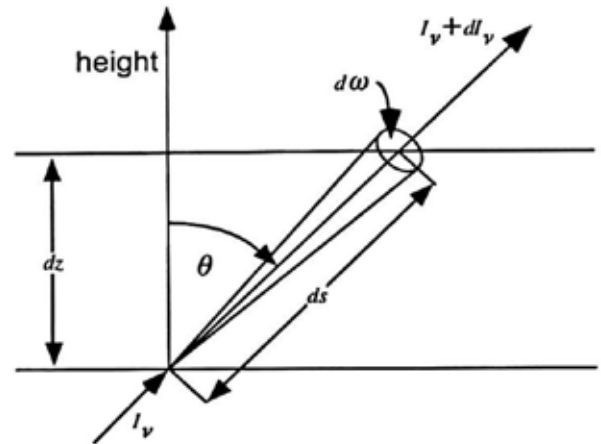
$s$ : coordinate along the ray

Optical depth

$$\tau = \int_z^\infty k_a dz'$$

Equation for radiance with the zenith angle of  $\theta$  ( $\mu = \cos\theta$ )

$$\mu \frac{dI}{d\tau} = I - B$$



Catling & Kasting (2017)

Upward radiance at the top of the atmosphere

$$I = B(T_s) \exp(-\tau_s) + \int_0^{\tau_s} B(T(\tau)) \exp(-\tau) d\tau \quad \text{Contribution function}$$

$$= B(T_s) \exp(-\tau_s) + \int_0^\infty B(T(z)) \boxed{k_a(z) \exp(-\tau(z))} dz$$

From surface

From atmosphere

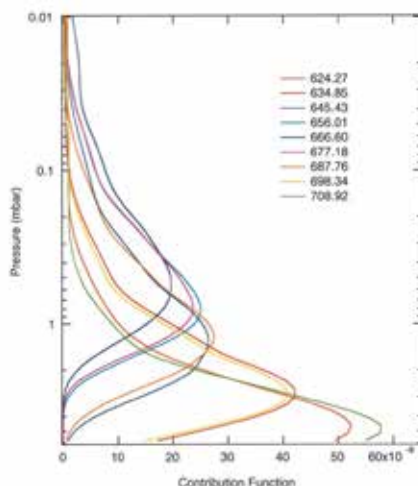
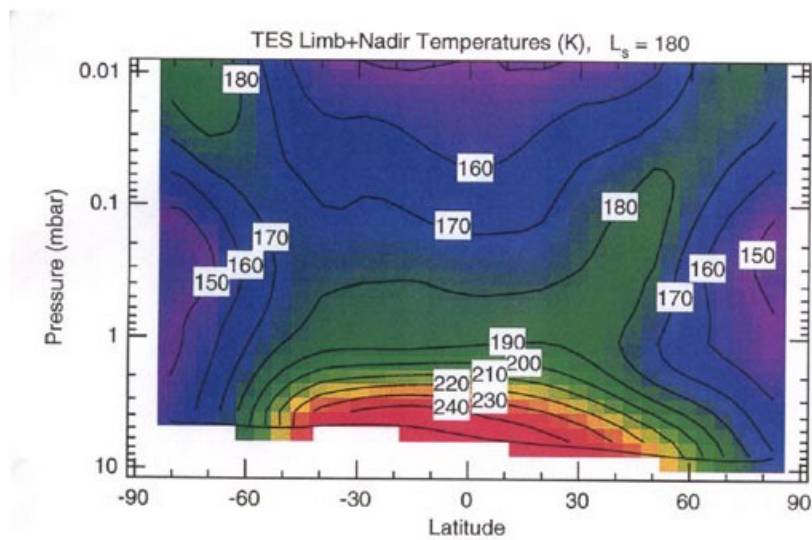


Figure 1. Contribution functions (functional derivatives of radiance with respect to temperature) for the wavenumbers used in the temperature retrievals. These functions were calculated using (10) and are for the nominal Thermal Emission Spectrometer (TES) 10 cm<sup>-1</sup> resolution. Units for the contribution functions are W cm<sup>-2</sup> sr<sup>-1</sup> cm<sup>-1</sup> K<sup>-1</sup>.

Contribution functions for Mars atmosphere in infrared (Conrath et al. 2000)

Temperature distribution retrieved from Mars Global Surveyor TES spectra (Smith et al. 2001)



Contribution functions for Venus atmosphere (Schofield & Taylor 1983)

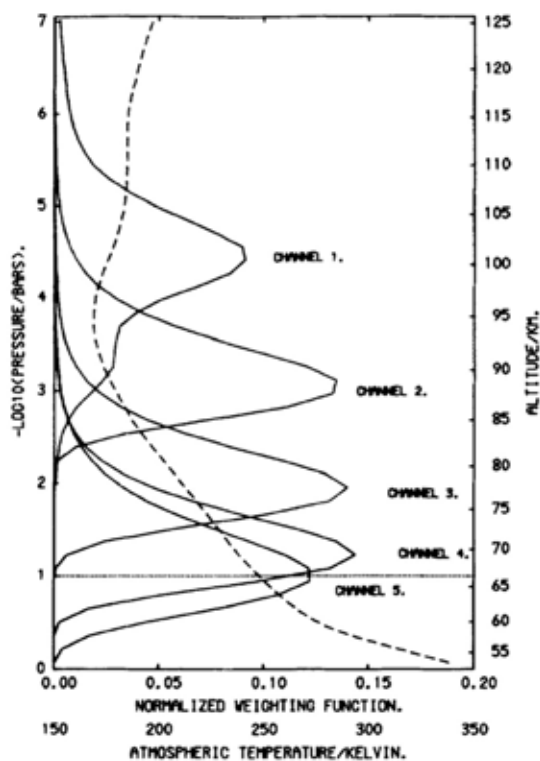


TABLE 1. OIR TEMPERATURE-SOUNDING CHANNELS - OPTICAL PROPERTIES

Channel*	Field of view** full angle (degrees)	Effective wavelength†		Spectral resolution†† ( $\text{cm}^{-1}$ )
		( $\text{cm}^{-1}$ )	( $\mu\text{m}$ )	
1	5.0	667.0	15.0	0.005
2	1.25	679.4	14.7	10.7
3	1.25	727.2	13.8	12.0
4	1.25	764.4	13.1	14.3
5	1.25	872.0	11.5	22.3



Obtained temperature distribution of Venus atmosphere (Schofield & Taylor 1983)

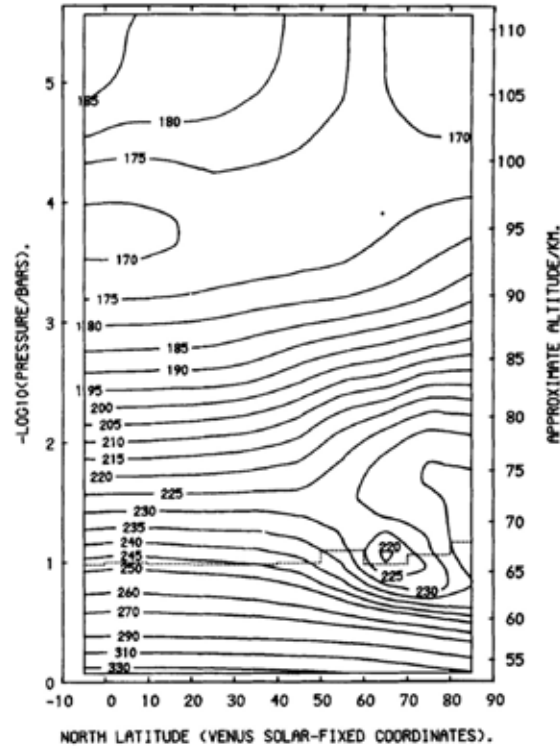


Figure 9. The retrieved zonal-mean temperature field and cloud structure. The temperature field is contoured as a function of pressure and latitude, and the altitude scale is averaged over latitude. Cloud unit optical depth at  $11.5\mu\text{m}$  is indicated by a dotted line.

Upward/downward flux ( $\theta' = \pi - \theta$ ) ( $\mu = \cos\theta$ )

$$F^\uparrow = \int_0^{2\pi} d\phi \int_0^{\pi/2} I(\theta) \cos\theta \sin\theta d\theta = 2\pi \int_0^1 I(\mu) \mu d\mu = \pi \int_0^1 I(\mu) \mu d\mu / \int_0^1 \mu d\mu$$

$$F^\downarrow = \int_0^{2\pi} d\phi \int_0^{\pi/2} I(\theta') \cos\theta' \sin\theta' d\theta' = -2\pi \int_{-1}^0 I(\mu) \mu d\mu = \pi \int_{-1}^0 I(\mu) \mu d\mu / \int_{-1}^0 \mu d\mu$$

Two-stream approximation :

$$F^\uparrow = \pi I(\bar{\mu})$$

$$F^\downarrow = \pi I(-\bar{\mu})$$

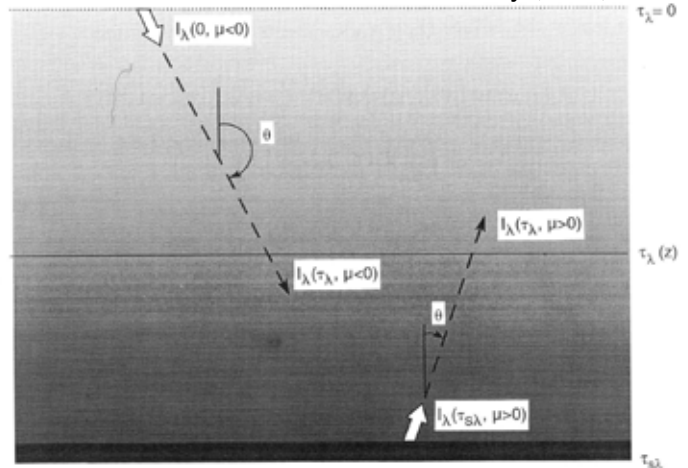
$$\tau^* = \bar{\mu}^{-1} \tau$$

$$B^* = \pi B$$

We have

$$\begin{aligned} \frac{dF^\uparrow}{d\tau^*} &= F^\uparrow - B^* \\ -\frac{dF^\downarrow}{d\tau^*} &= F^\downarrow - B^* \end{aligned} \quad (3.1)$$

Salby (1996)



$\bar{\mu} \approx 3/5$ , corresponding to the zenith angle of  $53^\circ$ , is frequently adopted.

# Radiative equilibrium in gray atmosphere

- Absorption coefficient in infrared does not depend on the wavelength.
- The atmosphere is transparent for solar radiation (visible wavelength)
- Solar energy reaching the surface is converted to thermal emission.
- Plane-parallel atmosphere with the incoming solar flux of  $F_0$ :

$$F_0 = (1 - A) \frac{S}{4}$$

S : solar constant  
A : albedo

Substituting

$$F^{net} = F^{\uparrow} - F^{\downarrow} \quad : \text{net upward flux}$$

$$F^{sum} = F^{\uparrow} + F^{\downarrow} \quad : \text{total flux}$$

into (3.1), we obtain

$$\frac{dF^{net}}{d\tau^*} = F^{sum} - 2B^* \quad (3.2)$$

$$\frac{dF^{sum}}{d\tau^*} = F^{net} \quad (3.3)$$

$F^{net}$  is equal to the incoming solar flux  $F^0$  :

$$F^{net} = F^0 \quad (3.4) \quad : \text{Net flux is invariant}$$

From (3.2)(3.4)

$$F^{sum} = 2B^* \quad (3.5) \quad : \text{Total flux is determined by the local temperature}$$

From (3.3)(3.4)

$$F^{sum} = F^0 \tau^* + F^{sum}(\tau^* = 0) \quad (3.6)$$

From (3.5)(3.6)

$$B^*(\tau^*) = \frac{F^0}{2} \tau^* + \frac{F^{sum}(\tau^* = 0)}{2} \quad (3.7)$$

Since  $F^{\downarrow} = 0$  at the top of the atmosphere ( $\tau^* = 0$ )

$$F^{sum}(\tau^* = 0) = F^{net} = F^0 \quad (3.8)$$

From (3.7)(3.8)

$$B^*(\tau^*) = \frac{F^0}{2} (\tau^* + 1) \quad (3.9)$$

Considering  $B^* = \sigma T^4$ , the temperature increases with decreasing the altitude (greenhouse effect).

Large  $\tau$  can lead to high temperatures  $\rightarrow$  Venus' hot atmosphere ( $\tau \sim 2000$ )

Earth's moderate temperature ( $\tau \sim 1$ )

The temperature at the top of the atmosphere ( $\tau^*=0$ ) is

$$T = (F^0/2\sigma)^{1/4} \quad \text{: "skin temperature" = temperature of the stratosphere}$$

This value is lower than the effective temperature.

From (3.4)(3.5) and the definition of  $F^{net}$  and  $F^{sum}$ , we obtain

$$F^\downarrow = \frac{F^{sum} - F^{net}}{2} = B^* - \frac{F^0}{2} \quad (3.10)$$

Emission from the surface is equal to the sum of the solar flux reaching the surface and the downward emission from the atmosphere:

$$B^*(T_s) = F^0 + F^\downarrow(\tau_s^*) \quad (3.11) \quad \begin{array}{l} T_s: \text{surface temperature} \\ \tau_s: \text{optical depth at the surface} \end{array}$$

From (3.10)(3.11), we obtain

$$\underbrace{B^*(T_s)}_{\text{surface}} = \underbrace{B^*(\tau_s^*)}_{\text{bottom of atmosphere}} + \frac{F^0}{2}$$

→ Temperature discontinuity exists at the surface (Note that  $B^* = \sigma T^4$ )

## Temperature structure in a gray atmosphere

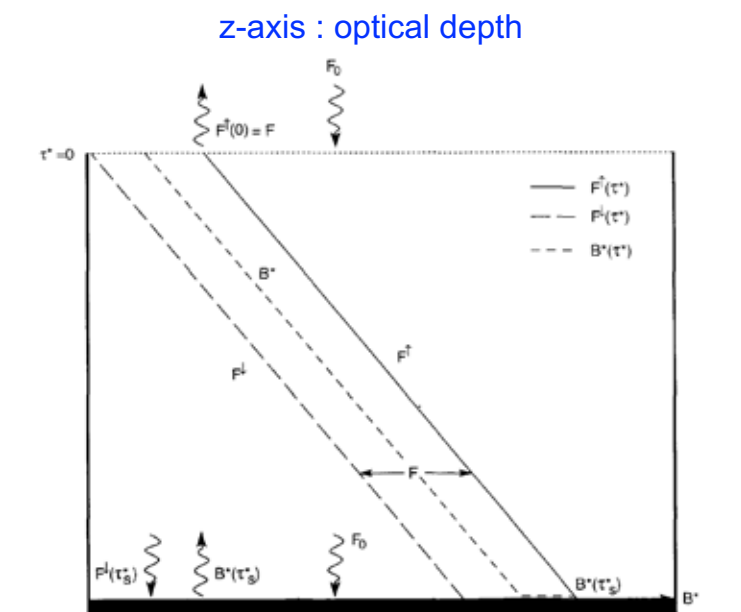


Figure 8.20 Upwelling and downwelling fluxes and emission in a gray atmosphere that is in radiative equilibrium with an incident SW flux  $F_0$  and a black underlying surface. Note: the emission profile is discontinuous at the surface.

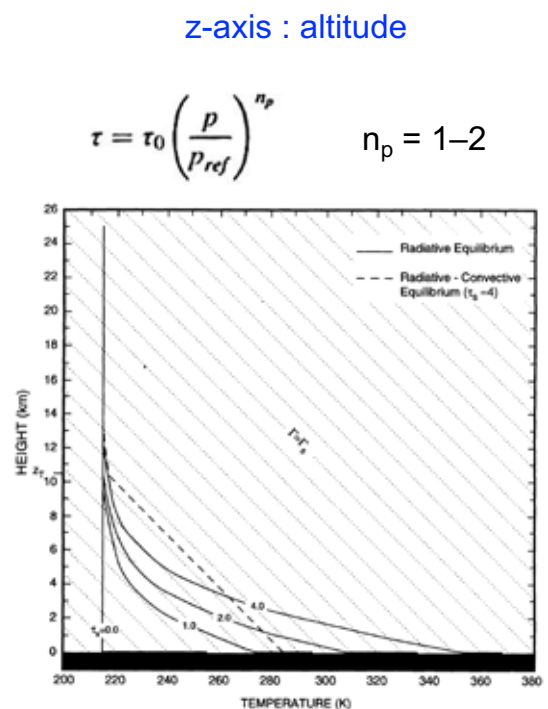


Figure 8.21 Radiative equilibrium temperature (solid lines) for the gray atmosphere in Fig. 8.20, with a profile of optical depth representative of water vapor (8.69), presented for several atmospheric optical depths  $\tau_s$ . Saturated adiabatic lapse rate (dotted lines) and radiative-convective equilibrium temperature for  $\tau_s = 4$  (dashed line) superposed.

# Radiative-convective equilibrium

The radiative equilibrium temperature profile can be unstable at low altitudes.

Assumptions for calculating convective adjustment:

- Vertical convection transports heat vertically, leading to an adiabatic lapse rate (troposphere).
- Above this convective region, the temperature profile remains to be the radiative equilibrium one (stratosphere).
- The surface temperature becomes equal to the atmospheric temperature at the bottom.
- The surface temperature is adjusted so that the upward energy flux

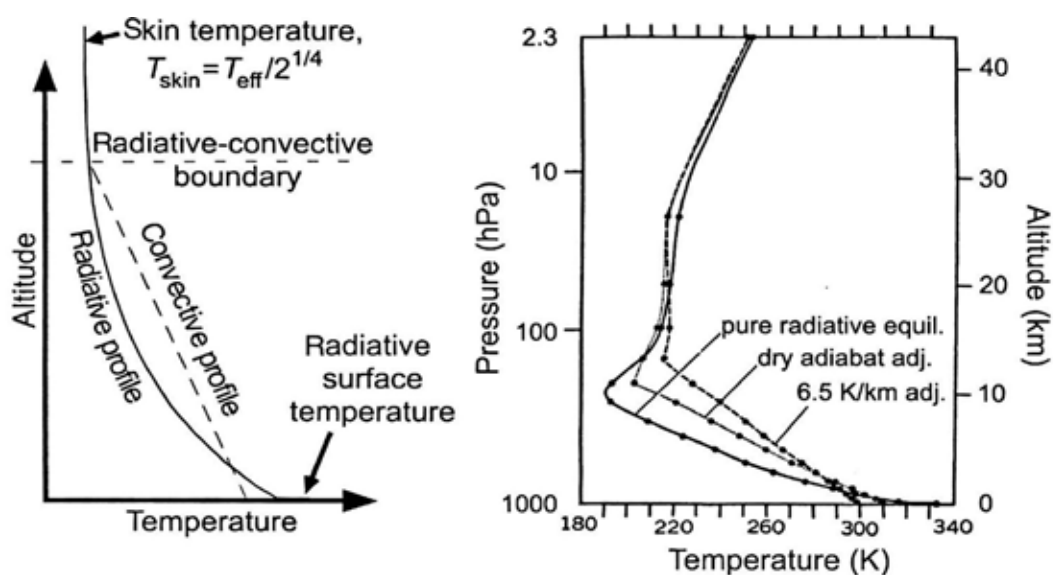
$$F^{\uparrow}(\tau^*) = B^*(T_s) \exp(\tau^* - \tau_s^*) - \int_{\tau_s^*}^{\tau^*} \exp(\tau^* - \tau') B^*(T(\tau')) d\tau'$$

becomes equal to the one for radiative equilibrium

$$F^{\uparrow}(\tau^*) = \frac{F^{sum} + F^{net}}{2} = \frac{F^0}{2}(\tau^* + 2)$$

at the top of the troposphere.

Radiative convective equilibrium  
for Earth's atmosphere  
(Manabe & Strickler 1964)



## Radiative-convective equilibrium solution for Venusian atmosphere (Pollack et al. 1980)

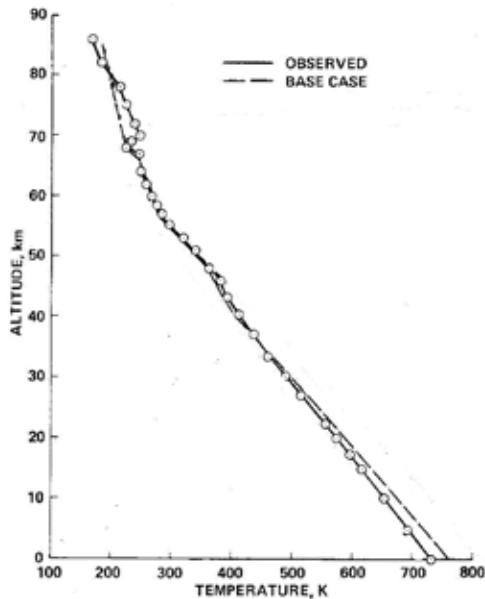


Fig. 2. Comparison between the observed temperature structure of Venus' lower atmosphere and that of several models, which are described in the main text.

## Net downward solar flux (Moroz et al. 1985)

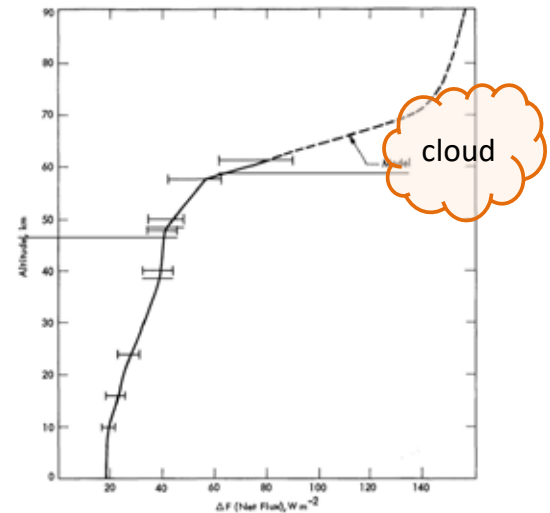


Figure 6-13. Globally Averaged Model of Total Solar Flux

## Radiative-convective equilibrium solution for Martian atmosphere

without dust

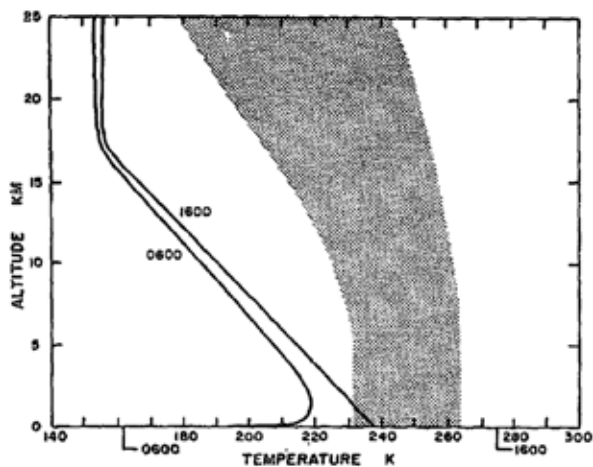


FIG. 1. Martian temperature calculations. The stippled area represents temperatures reported by Kliore *et al.* (1972) and Hanel *et al.* (1972). The lines are theoretical profiles for a pure CO<sub>2</sub> atmosphere, at 1600 and at 0600 hours (the coldest time). Both theory and observation refer to mid-latitude summer conditions. The tags indicate the ground temperatures. In the case of the 1600 theoretical profile a strong boundary layer is indicated.

with dust

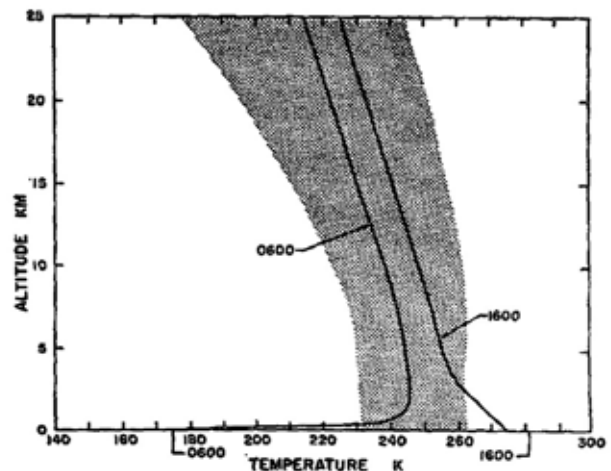
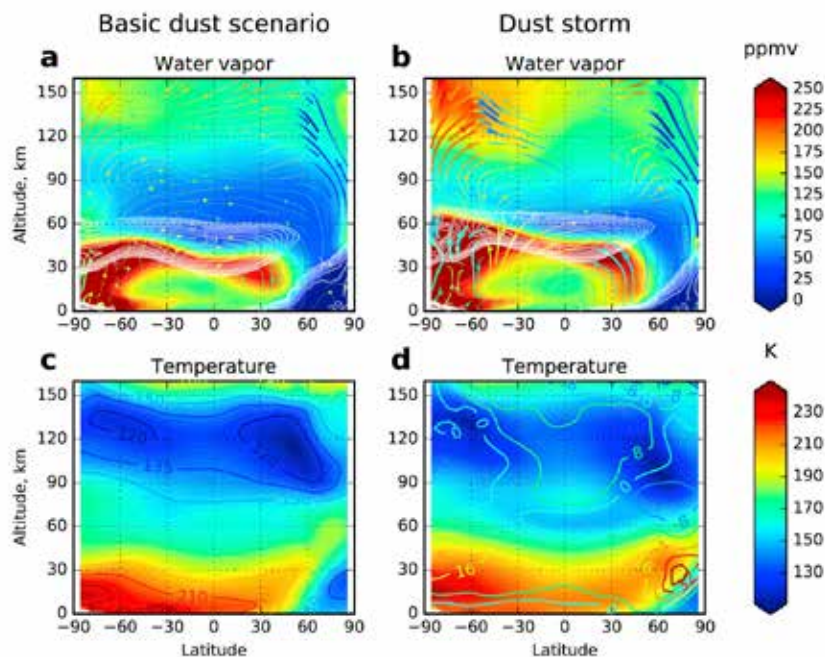


FIG. 2. Same as Fig. 1 except that the atmosphere contains an extra solar absorber, evenly mixed with the atmosphere at all levels, and having an optical depth of 0.10 at all wavelengths. Note the weak boundary layer at 1600.

Gierasch & Goody (1972)

## Meridional cross sections in Mars GCM



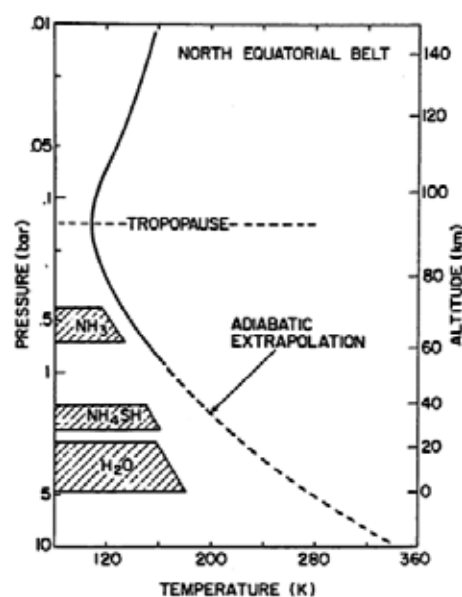
Shaposhnikov et al. (2019)

**Figure 2.** Latitude-altitude cross-sections of the quantities simulated for the “basic” dust scenario (left column) and the MY28 dust storm (right column): (a) water vapor (shaded), water ice (white contours), and the meridional flux of water vapor (the lines with arrows, the color and thickness of which indicate the vertical direction and magnitude, correspondingly); (b) is the same as in panel (a), but for the dust storm of MY28; (c) temperature (shaded) for the “basic” dust scenario; (d) is the same as in (c), but for the MY28 dust storm scenario, except for the contour lines that show the temperature difference between (d) and (c). All fields are averaged zonally and over the period between  $L_s = 250^\circ$  and  $270^\circ$ .

## Internal heat of gas giants

Internal heat sources include the rainout of helium-rich droplets and, possibly, continued Kelvin-Helmholtz contraction.

The Kelvin–Helmholtz mechanism is an astronomical process that occurs when the surface of a star or a planet cools. The cooling causes the pressure to drop and the star or planet shrinks as a result.



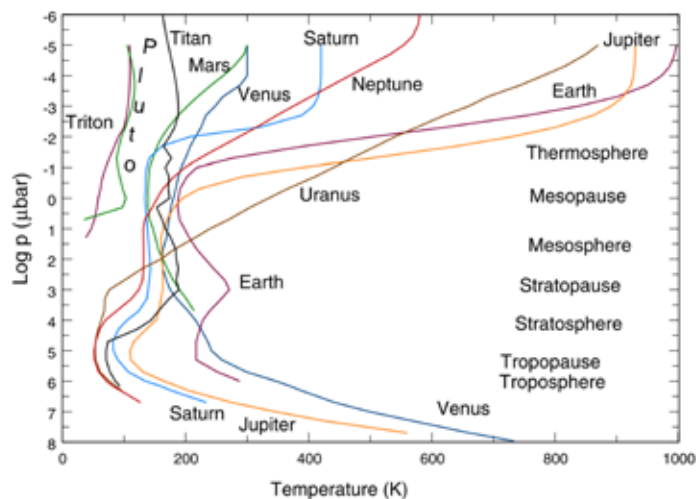
**Fig. 1.26** Jovian temperatures and schematic cloud structure, based primarily on Voyager infrared data. The height scale starts at the cloud base, approximately 5 bars. [After KUNDE et al. (1982).]

Chamberlain & Hunten (1987)

**TABLE 1.3** *Characteristics of the Jovian Planets*

	Jupiter	Saturn	Uranus	Neptune
Mean density (g/cm <sup>3</sup> )	1.34	0.70	1.58	2.30
Effective temperature (°K)	124.4	95.0	58	55.5
Equilibrium temperature (°K)	109.5	82.3	57	46
Total flux/solar heat	1.668	1.78	<1.3	1.1
Internal flux (erg cm <sup>-2</sup> sec <sup>-1</sup> )	5444	2000	<180	285
Adiabatic lapse rate (°K/km)	1.9	0.84	0.85	0.86
Tropopause temperature (°K)	105	85	54	52
Tropopause pressure (mbar)	140	80	100	200
Exospheric temperature (°K)	700–1000	420	700	—

- Outward thermal flux > Incoming solar flux (Jupiter, Saturn)
- The Earth's internal heat flux is 1/40000 of the Incoming solar flux. This heat comes from a combination of residual heat from planetary accretion and heat produced through radioactive decay.



(Mueller-Wodarg et al.)

**Table 1.3** Thermodynamic properties and lapse rates on various planets.

Body	Main gases	Molar mass, $M$ g mol <sup>-1</sup>	$\bar{R}$ J kg <sup>-1</sup> K <sup>-1</sup>	$g$ m s <sup>-2</sup>	$c_p$ J kg <sup>-1</sup> K <sup>-1</sup>	$\Gamma_a = g/c_p$ troposphere K km <sup>-1</sup>	$\Gamma$ observed in troposphere K km <sup>-1</sup>
Venus	CO <sub>2</sub> , N <sub>2</sub>	43.45	189	8.901	930	9.5	~8.0
Earth	N <sub>2</sub> , O <sub>2</sub>	28.97	287	9.81	1004	9.8	~6.5
Mars	CO <sub>2</sub> , N <sub>2</sub>	43.5	191	3.72	850	4.4	~2.5
Jupiter	H <sub>2</sub> , He	2.22	3745	24.25	10 988	2.1–2.45*	1.9
Saturn	H <sub>2</sub> , He	2.14	3892	10.0	10 658	0.7–1.1	0.85
Titan	N <sub>2</sub> , CH <sub>4</sub>	28.67	290	1.36	1044	1.3	1.0–1.4
Uranus	H <sub>2</sub> , He	2.3	3615	8.80	8643	0.7–1.1	0.75
Neptune	H <sub>2</sub> , He	2.3	3615	11.1	8643	0.85–1.34	0.95
HD209458b	H <sub>2</sub>	2.0	4160	18.5	14 300	1.3	~0.2 (model)**

\* On giant planets,  $g$  varies from equator to pole.

\*\* Menou and Rauscher (2009).

Catling & Kasting (2017)



# Adiabatic lapse rate in log-pressure coordinate

Why is the tropospheric lapse rate in the log-pressure coordinate similar between the planets?

From

$$\frac{dT}{dz} = -\frac{g}{C_p}$$

and

$$\frac{dp}{dz} = -\frac{gp}{RT},$$

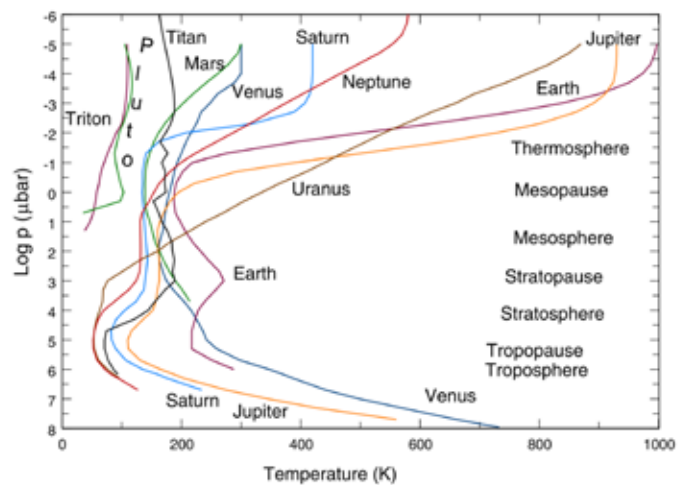
we get

$$\frac{dT}{d(\ln p)} = \frac{R}{C_p} T$$

Here

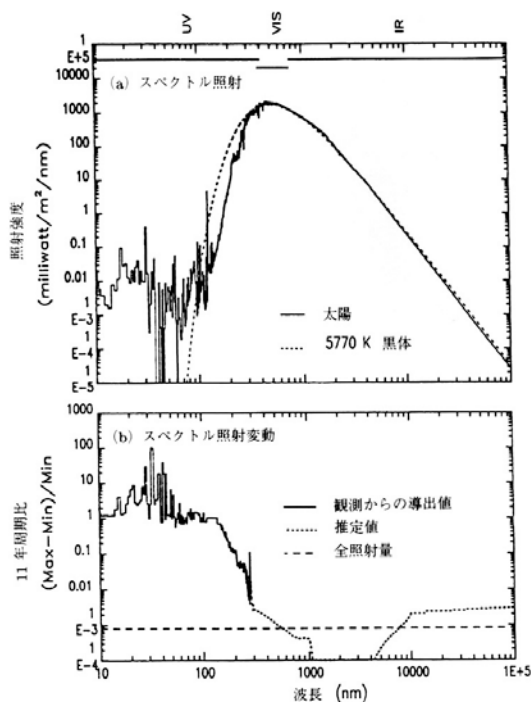
$$\frac{C_p}{R} = \frac{\text{degree of freedom}}{2} + 1$$

Degree of freedom = 5 at room temperature for a linear molecule (H<sub>2</sub>, O<sub>2</sub>, N<sub>2</sub>, CO<sub>2</sub>, ..)



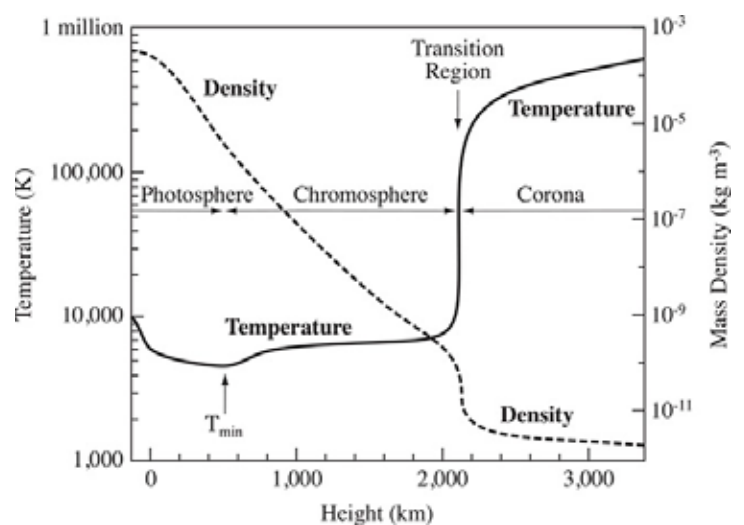
## Heating of the upper atmosphere

Solar spectrum



(ブレッケ, 超高層大気物理学)

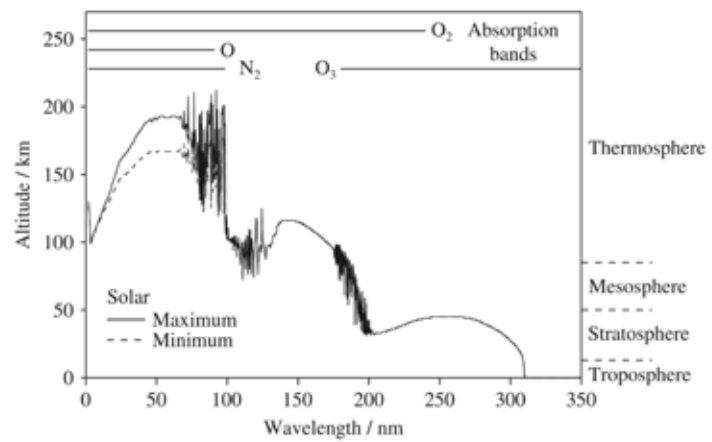
Structure of the solar atmosphere



(Eugene Avrett, Smithsonian Astrophysical Observatory)



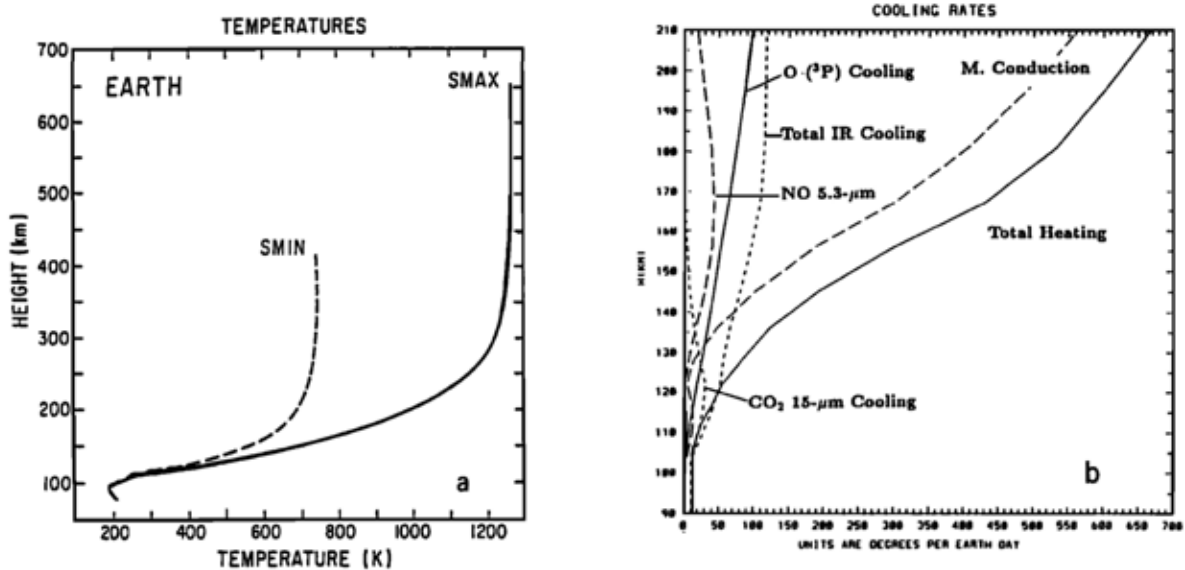
chromosphere



The altitude of unit optical depth for vertical solar radiation. The principal absorption bands are shown. Adapted from Meier (1991); an early version of this figure appeared in Herzberg (1965). Figure courtesy of Dr J. Lean and Dr R. Meier.

(Andrews 2010)

## Energy balance of the thermosphere

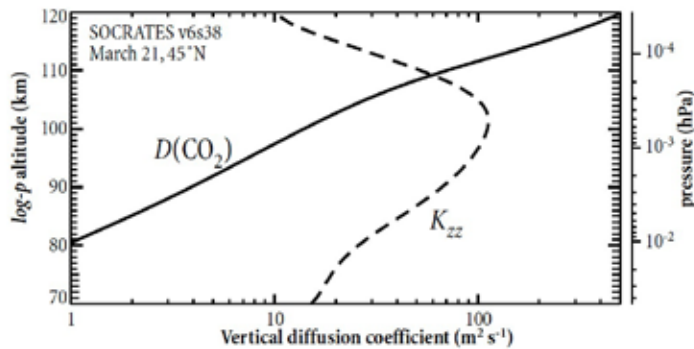


Bougher et al. (1994)

Molecular diffusion coefficient (for CO<sub>2</sub>) (Chapman and Cowling 1970)

$$K = 1.38 \times 10^5 \cdot \frac{1}{\rho} \cdot \left( \frac{T}{273} \right)^{0.933}$$

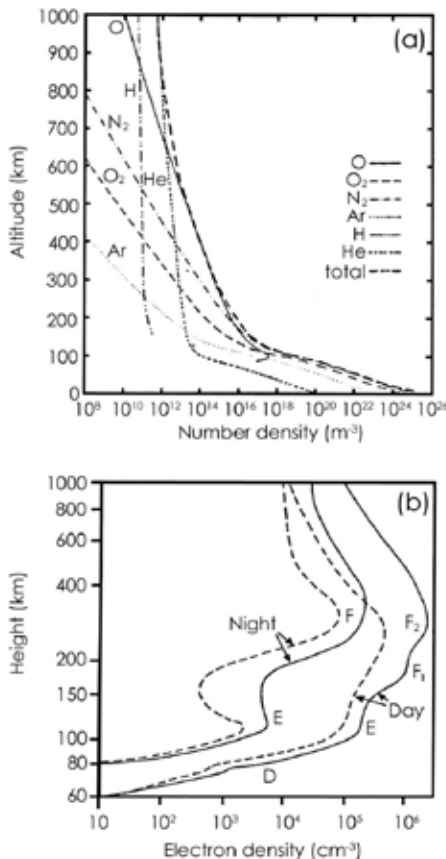
$\rho$  : Atmospheric density  
 $T$  : Atmospheric temperature



Chabrillat et al. (2002)

**Figure 1.** Vertical profiles of the eddy diffusion coefficient and the CO<sub>2</sub> molecular diffusion coefficient, using the SOCRATES baseline model. Latitude 45°North, equinox (March 21), solar minimum conditions.

## Composition of Earth's upper atmosphere



Borderick, 2010

## Homopause levels

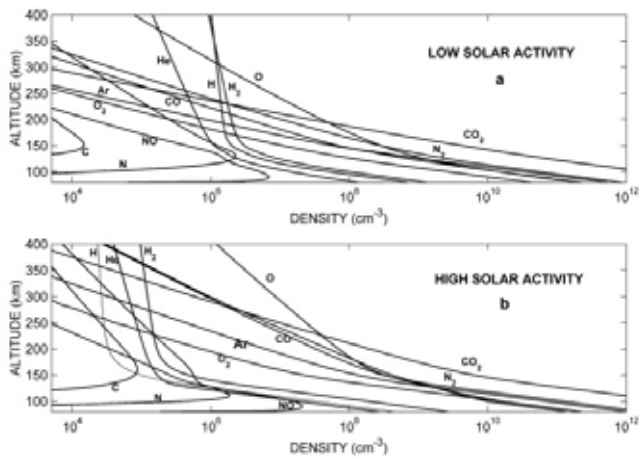
**Table 1.1** Homopause levels. (Sources: Atreya et al. (1991), p. 145; Atreya et al. (1999).)

Planet	Altitude (km)	Pressure ( $\mu$ bar)	Number density (molecules $\text{cm}^{-3}$ )
Venus	130–135	0.02	$7.5 \times 10^{11}$
Earth	~100	0.3	$10^{13}$
Mars	~130	0.002	$10^{10}$
Jupiter	~385 above 1 bar level	1	$1.4 \times 10^{13}$
Saturn	~1140 above 1 bar level	0.005	$1.2 \times 10^{11}$
Titan	800–850	~0.0006	$2.7 \times 10^{10}$
Uranus	~354–390 above the 1 bar level	~20–40	$1\text{--}2 \times 10^{15}$
Neptune	~586–610 above the 1 bar level	~0.02	$10^{13}$

Catling & Kasting (2017)

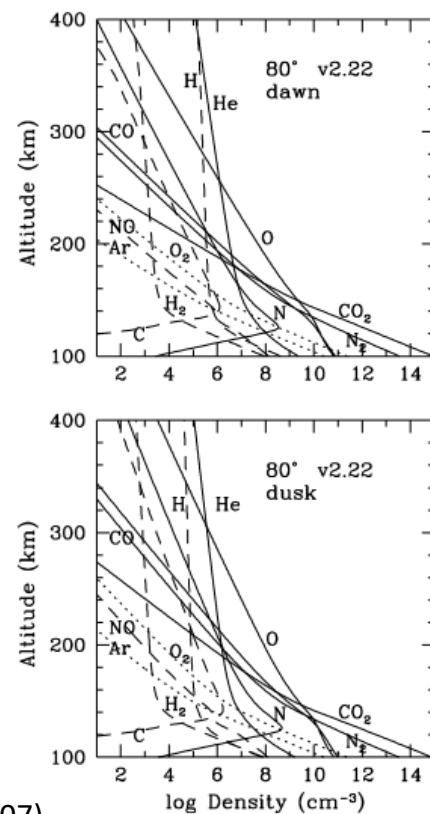
## Mars and Venus thermospheres

### Mars



(Bakalian et al. 2006)

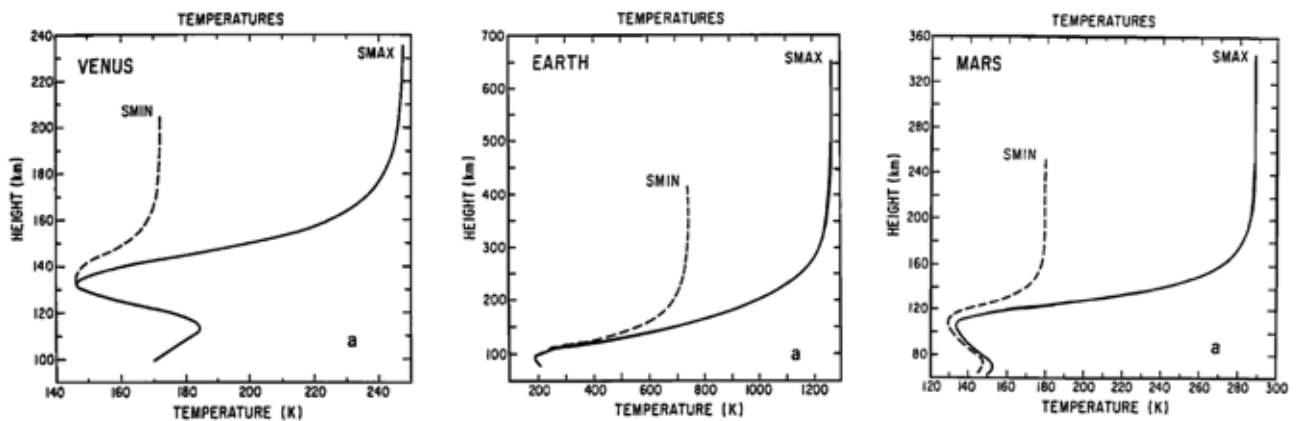
### Venus



(Fox & Kasprzak, 2007)

## Thermospheres of the terrestrial planets

(Bougher et al. 1994)



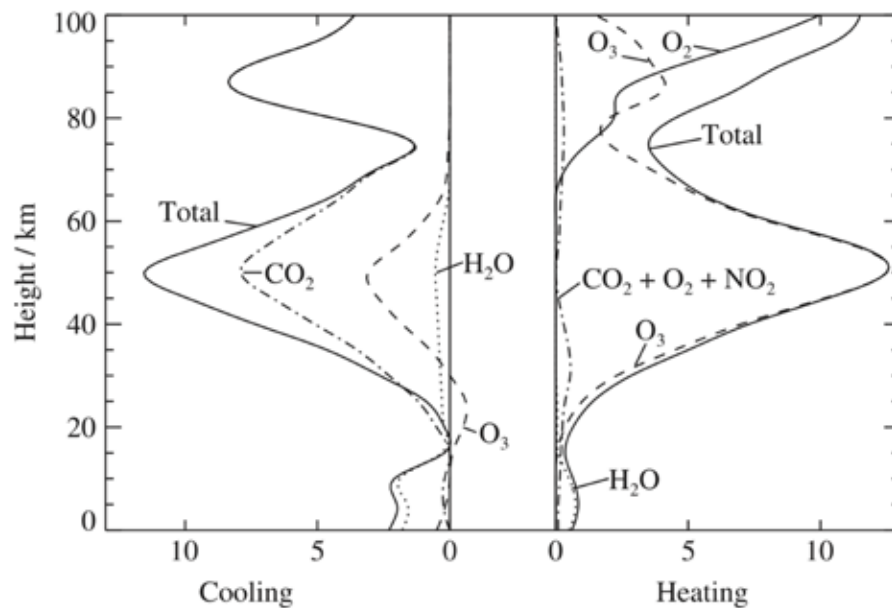
Thermospheric temperature

Earth: 1000 K Venus: 250 K Mars: 270 K

The coolants in Earth's thermosphere,  $\text{CO}_2$  and  $\text{NO}$ , are relatively ineffective because of their low concentrations.

On Venus and Mars, the atmospheres are almost  $\text{CO}_2$ , which makes the upper atmospheres cold through efficient radiative cooling.

## Radiative energy balance at each altitude



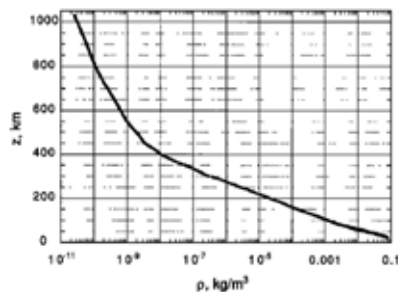
Global-mean vertical profiles of the short-wave heating rate and the long-wave cooling rate, in  $\text{K day}^{-1}$ , including contributions from individual gases. Adapted from [London \(1980\)](#).

(Andrews 2010)

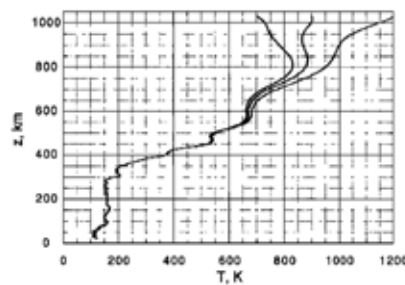
## Jovian thermospheric temperature

22,876

SEIFF ET AL.: THERMAL STRUCTURE OF JUPITER'S ATMOSPHERE



**Figure 26.** Density of the upper atmosphere as a function of altitude, derived from measured probe decelerations. A major change in density scale height occurred between 350 and 550 km. This coincides with the onset of diffusive separation. The steeper slope above 550 km indicates a major warming of the upper atmosphere.



**Figure 28.** Temperature structure of the upper atmosphere calculated from measured densities, derived pressures, and the mean molecular weight profile of Figure 19. The effect of three widely differing temperature assumptions at the initial altitude is shown. The three profiles effectively converge at 750 km altitude. Waves in the thermal structure and the deep isothermal layer below 300 km are conspicuous.



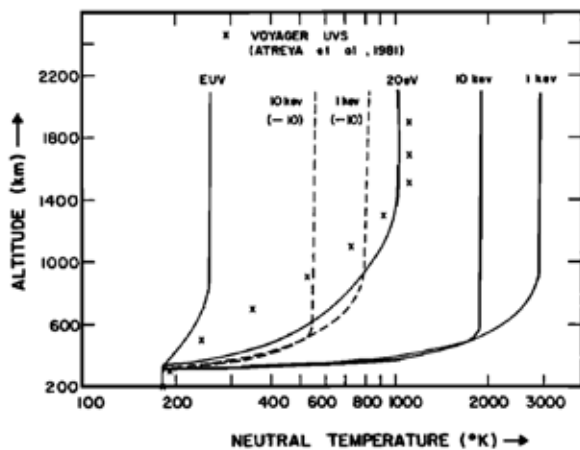


Fig. 16. Neutral temperature as a function of altitude for several cases of interest. The EUV results use only photoelectrons as a heat source. The 20-eV case considers the heating due to 20-eV electrons with an energy flux equal to  $0.5 \text{ ergs cm}^{-2} \text{ s}^{-1}$ . The 1- and 10-keV auroral electron cases show the effects of electron heating from 1- and 10-keV electrons with an energy flux of  $10 \text{ ergs cm}^{-2} \text{ s}^{-1}$  and for auroral heating rates diluted by a factor of 10 to illustrate the possible global effects of auroral heating. The Voyager UVS stellar occultation-derived profile is shown by the crosses.

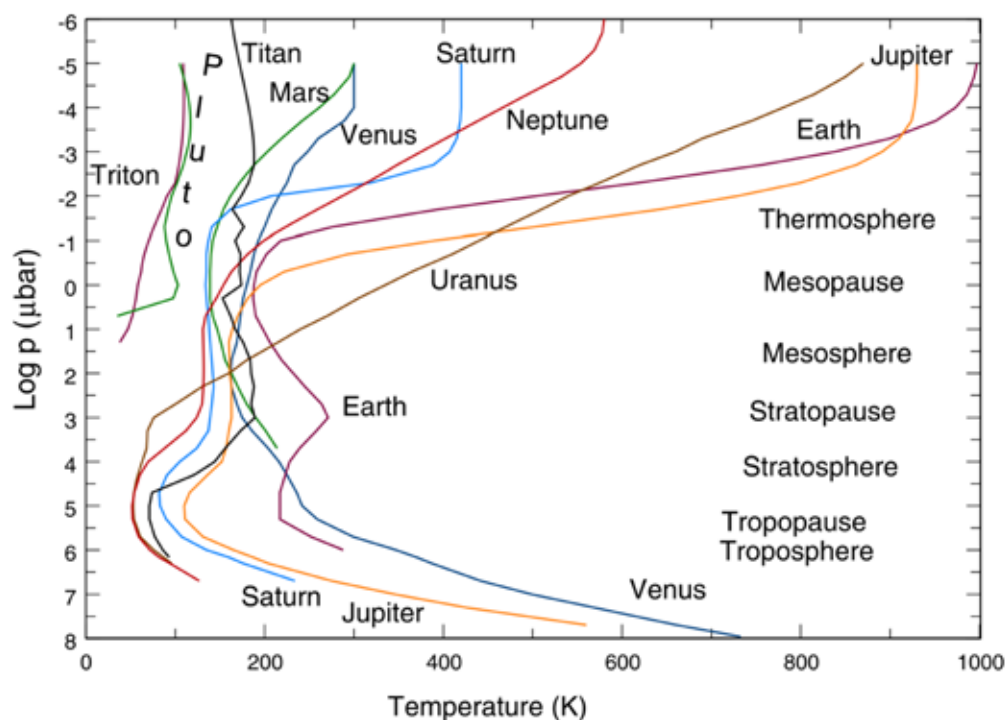
The temperature rise across the thermosphere due to solar UV heating is predicted to be  $<100 \text{ K}$ .

A much stronger source of heat must be present.

- Precipitating electrons ?
- Wave heating (gravity wave, acoustic wave) ?

Waite et al. (1983)

## Vertical structures of planetary atmospheres



(Mueller-Wodarg et al.)