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

$$-S\Delta p = g\rho S\Delta z$$

$$\therefore \frac{dp}{dz} = -g\rho \qquad (1.1)$$

p:pressure *z*: altitude ρ :mass density(kg/m³)

This is equivalent to the vertical momentum equation

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



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:

$$p = \rho RT \tag{1.2}$$

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

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

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

 $p_{\rm s}$: 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:

 $dH = c_{y}dT + pd\alpha \qquad (2.1)$

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

$$pd\alpha + \alpha dp = RdT \tag{2.2}$$

Combining (2.1)(2.2) yields

$$dH = c_p dT - \alpha dp \tag{2.3}$$

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

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

$$c_p dT = \frac{RT}{p} dp$$
$$c_p d(\ln T) = Rd(\ln p)$$
$$\therefore T = const. \times p^{R/C_p}$$

Then, the potential temperature $\boldsymbol{\theta}$ defined as

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

is conserved in adiabatic processes.

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

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

 $\Gamma_d = g/C_p$: 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 ⁻¹	\overline{R} J kg ⁻¹ K ⁻¹	g m s ⁻²	<i>с_р</i> Ј kg ⁻¹ К ⁻¹	$\Gamma_{\rm a} = g/c_p$ troposphere K km ⁻¹	Γ observed in troposphere K km ⁻¹
Venus	CO ₂ , N ₂	43.45	189	8.901	930	9.5	~8.0
Earth	N ₂ , O ₂	28.97	287	9.81	1004	9.8	~6.5
Mars	CO ₂ , N ₂	43.5	191	3.72	850	4.4	~2.5
Jupiter	H ₂ , He	2.22	3745	24.25	10 988	2.1-2.45*	1.9
Saturn	H ₂ , He	2.14	3892	10.0	10 658	0.7-1.1	0.85
Titan	N ₂ , CH ₄	28.67	290	1.36	1044	1.3	1.0-1.4
Uranus	H ₂ , He	2.3	3615	8.80	8643	0.7-1.1	0.75
Neptune	H ₂ , He	2.3	3615	11.1	8643	0.85-1.34	0.95
HD209458b	H ₂	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

$$\left(\rho_{\rm p}, T_{\rm p} \right) \bar{\rho}, \bar{T}$$

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

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

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

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

$$\overline{T} = T_0 - \Gamma z$$

$$T_p = T_0 - \Gamma_d z$$

$$\Gamma = - d\overline{T}/dz \quad \text{: ambient lapse rate}$$

$$\Gamma_d = g/C_p \quad \text{: adiabatic lapse lapse rate}$$

$$T_0 \quad \overline{T}$$

From (2.6)(2.7)

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

When $\Gamma_d - \Gamma$ is positive, an oscillating solution exists. The buoyancy frequency *N* is given by

$$N^{2} = g \frac{\Gamma_{d} - \Gamma}{T_{0}} = g \frac{\partial \ln \overline{\theta}}{\partial z}$$

Static stability is defined by

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

Three types of static stability contitions:

$$\begin{array}{ll} \Gamma_{d} - \Gamma \geq 0 & \leftrightarrow \ S \geq 0 & \leftrightarrow \ \partial \bar{\theta} / \partial z \geq 0 & : \ \text{stable} \\ \Gamma_{d} - \Gamma \equiv 0 & \leftrightarrow \ S \equiv 0 & \leftrightarrow \ \partial \bar{\theta} / \partial z \equiv 0 & : \ \text{neutral} \\ \Gamma_{d} - \Gamma \leq 0 & \leftrightarrow \ S \leq 0 & \leftrightarrow \ \partial \bar{\theta} / \partial z \leq 0 & : \ \text{unstable} \\ & \Gamma = - \ d \bar{T} / dz & : \ \text{ambient lapse rate} \\ & \Gamma_{d} = g / C_{p} & : \ \text{adiabatic lapse lapse rate} \end{array}$$



Stability of Earth's atmosphere



Stability of Venusian atmosphere







Akatsuki radio occultation (Imamura et al. 2017)

Stability of Martian atmosphere



Potential Temperature, K

Radiation

Energy balance of a planet

Inflow : Visible wavelength radiation from Sun

Outflow : Infrared radiation from the surface and the atmosphere



Logarithm of the black-body spectral radiance $B_{\lambda}(I)$, plotted against the logarithm of wavelength λ , for I = 6000 K, a typical temperature of the solar photosphere, and 288 K, the Earth's mean surface temperature.

Planck function (J/m2/s/str/Hz)

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

h: Planck's constant c: speed of light v: frequency k: Boltzmann's constant



Integration for wavelength and for solid angle over a hemisphere

$$\frac{\int_{0}^{2\pi} d\phi \int_{0}^{\pi/2} d\theta \sin \theta \cos \theta}{\int_{0}^{\infty} d\nu B_{\nu}(T)} = \pi \int_{0}^{\infty} d\nu B_{\nu}(T) = \sigma T^{4}$$

Integration for solid angle

Integration for frequency

 σ : Stefan-Boltzmann's constant

Equilibrium between solar radiation and planetary infrared radiation

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

A: albedo (0.3 for Earth) a:planetary radius S: solar constant (1370 W m⁻² for Earth)

For Earth, T=255 K : "effective temperature"

Interaction between electromagnetic waves and molecules



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)²⁶⁾

柴田 (1999)





柴田 (1999)

Catling & Kasting (2017)

Vibrational energy levels

$$E_{v} = hv_0 \left(v + \frac{1}{2} \right)$$

v: vibrational quantum number

Rotational energy levels

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

J : rotational quantum number



Figure 2.23 The meaning of the P, Q, and R branches in vibrational-rotational transitions, showing a vibrational transition Δv and superposed rotational transitions, ΔJ . 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 ΔJ transitions.

Catling & Kasting (2017)



Chamberlain & Hunten (1987)

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).]

Radiative transfer in plane-parallel atmosphere

Radiative transfer equation

$$dI = (absorption) + (emission) = -k_a Ids + k_a B(T)ds$$

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

$$\lim_{k_a: absorption coefficient}$$

s: coordinate along the ray

Optical depth

$$\tau = \int_{z}^{\infty} k_{a} dz'$$

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

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



Upward radiance at the top of the atmosphere

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

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

From surface

From atmosphere



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

Plate 1. Contribution functions (functional derivatives of radiance with respect to temperature) for the www.ensurbers used in the temperature retrievals. These functions were calculated using (10) and are for the nominal Thermal Illumians Spectrometer (TES) 10 cm⁻¹ resolution. Units for the contribution functions are W cm⁻² r^{-1} cm⁻¹ K^{-1} . Temperature distribution retrieved from Marg Global Surveyor TES spectra (Smith et al. 2001)



Contribution functions for Venus atmosphere (Schofield & Taylor 1983)



TARIE 1 OIR TEMPERATURE-SOUNDING CHANNELS - OPTICAL PROPER	THE

Channel*	Field of view** full angle	Effective wa	Spectral resolution ^{††}	
	(degrees)	(cm - 1)	(µm)	(cm ⁻¹)
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 $11-5\,\mu m$ is indicated by a dotted line.

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

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

Two-stream approximation :

$$F^{\uparrow} = \pi I(\overline{\mu})$$
$$F^{\downarrow} = \pi I(-\overline{\mu})$$
$$\tau^* = \overline{\mu}^{-1}\tau$$
$$B^* = \pi B$$

We have

$$\frac{dF^{\uparrow}}{d\tau^{*}} = F^{\uparrow} - B^{*}$$

$$-\frac{dF^{\downarrow}}{d\tau^{*}} = F^{\downarrow} - B^{*}$$
(3.1)



 $\overline{\mu} \approx 3/5$, corresponding to the zenith angle of 53°, 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 F₀:

$$F_0 = (1 - A) \frac{S}{4}$$
 S : solar constant
A : albedo

Substituting

 $F^{net} = F^{\uparrow} - F^{\downarrow}$: net upward flux $F^{sum} = F^{\uparrow} + F^{\downarrow}$: total flux

into (3.1), we obtain

$$\frac{dF^{net}}{d\tau^*} = F^{sum} - 2B^* \qquad (3.2)$$
$$\frac{dF^{sum}}{d\tau^*} = F^{net} \qquad (3.3)$$

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

$$F^{net} = F^0 \tag{3.4}$$

From (3.2)(3.4)

$$F^{sum} = 2B^* \tag{3.5}$$

From (3.3)(3.4)

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

From (3.5)(3.6)

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

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

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

From (3.7)(3.8)

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

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

Large τ can lead to high temperatures \rightarrow Venus' hot atmosphere ($\tau \sim 2000$) Earth's moderate temperature ($\tau \sim 1$)

- 4) : Net flux is invariant
 - : Total flux is determined by the local temperature

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

$$T = (F^0/2\sigma)^{1/4}$$
 : "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

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

$$B^*(T_s) = F^0 + F^{\downarrow}(\tau_s^*)$$

(3.11) T_s : surface temperature τ_s : optical depth at the surface

From (3.10)(3.11), we obtain

$$B^{*}(T_{s}) = B^{*}(\tau_{s}^{*}) + \frac{F^{0}}{2}$$

surface bottom of
atmosphere

→ 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.

z-axis = altitude

$$\tau = \tau_0 \left(\frac{p}{p_{ref}}\right)^{n_p} \qquad n_p = 1 - 2$$



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 τ_r . Saturated adiabatic lapse rate (dotted lines) and radiative-convective equilibrium temperature for $\tau_r = 4$ (dashed line) superposed.

Salby (1996)

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)



Catling & Kasting (2017)

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



Radiative-convective equilibrium solution for Martian atmosphere



F10. 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 CO₂ 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°.

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)

	Jupiter	Saturn	Uranus	Neptune
Mean density (g/cm ³)	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 ⁻² sec ⁻¹)	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	_

TABLE 1.3 Characteristics of the Jovian Planets

- 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.



Table 1.3 Thermodynamic properties and lapse rates on various planets.

Body	Main gases	Molar mass, M g mol ⁻¹	R J kg ^{−1} K ^{−1}	g m s ⁻²	c _p J kg ⁻¹ K ⁻¹	$\Gamma_n = g/c_p$ troposphere K km ⁻¹	Γ observed in troposphere K km ⁻¹
Venus	CO2, N2	43.45	189	8.901	930	9.5	~8.0
Earth	N2, O2	28.97	287	9.81	1004	9.8	~6.5
Mars	CO2, N2	43.5	191	3.72	850	4.4	~2.5
Jupiter	H ₂ , He	2.22	3745	24.25	10 988	2.1-2.45*	1.9
Saturn	H ₂ , He	2.14	3892	10.0	10 658	0.7-1.1	0.85
Titan	N2, CH4	28.67	290	1.36	1044	1.3	1.0-1.4
Uranus	H ₂ , He	2.3	3615	8.80	8643	0.7-1.1	0.75
Neptune	H ₂ , He	2.3	3615	11.1	8643	0.85-1.34	0.95
HD209458b	H ₂	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).

Adiabatic lapse rate in log-pressure coordinate

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



Degree of freedom = 5 at room temperature for a linear molecule (H₂, O₂, N₂, CO₂, ..)

Heating of the upper atmosphere



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)



Figure 1. Vertical profiles of the eddy diffusion coefficient and the CO_2 molecular diffusion coefficient, using the SOCRATES baseline model. Latitude 45°North, equinox (March 21), solar minimum conditions.



Composition of Earth's upper atmosphere

Homopause levels

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

Planet	Altitude (km)	Pressure (µbar)	Number density (molecules cm ⁻³)
Venus	130-135	0.02	7.5 × 10 ¹¹
Earth	~100	0.3	1013
Mars	~130	0.002	1010
Jupiter	~385 above 1 bar level	1	1.4×10^{13}
Saturn	~1140 above 1 bar level	0.005	1.2×10^{11}
Titan	800-850	~0.0006	2.7×10^{10}
Uranus	~354-390 above the 1 bar level	~20-40	$1-2 \times 10^{15}$
Neptune	~586-610 above the 1 bar level	~0.02	1013

Catling & Kasting (2017)

Borderick, 2010



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, CO_2 and NO, are relatively ineffective because of their low concentrations.

On Venus and Mars, the atmospheres are almost 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 $K \, day^{-1}$, 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 $10 \, \text{keV}$ auroral electron cases show the effects of electron heating from 1- and 10-keV electrons with an energy flux of 10 ergs cm⁻² 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.

Waite et al. (1983)

The temperature rise across the thermosphere due to solar UV heating is predicted to be <100 K.

A much stronger source of heat must be present.

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



Vertical structures of planetary atmospheres

(Mueller-Wodarg et al.)