

ASTRONOMICAL, PHYSICAL, AND METEOROLOGICAL PARAMETERS  
FOR PLANETARY ATMOSPHERES

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*A newly compiled table of astronomical, physical, and meteorological parameters for planetary atmospheres is presented together with formulae and explanatory notes for their application and a complete listing of sources.*

Although the reconnaissance of the Solar System by spacecraft in the past fifteen years has virtually recalibrated the measure of planetary parameters, there is an evident lag in the comprehensive digestion of the new data into a form that can serve as a ready reference for comparative planetology. Certainly there is not yet available anything like a Solar System analogue to Allen's *Astrophysical Quantities* (1973) for the everyday use of the practicing space scientist. The tabular appendix to *The New Solar System*, "Planetary and Satellite Characteristics" (Beatty, O'Leary, and Chaiken, 1982) is a useful reference for orbital and physical parameters, though already partly out of date. Hubbard's (1984a) *Planetary Interiors* textbook contains a number of excellent tables for the comparison of internal structure, heat flow, and magnetic field characteristics of the planets and satellites. What is still lacking, however, is a succinct tabulation of specifically atmospheric and meteorological parameters, updated with the most recent Voyager measurements of the Jovian planets, together with the relevant astronomical and physical structure data.

The purpose of this appendix is to offer a partial remedy in the form of a single table designed specifically to provide a listing of the most important parameters for the comparative study of planetary atmospheres. It represents a compilation not only of such fundamental and well-measured quantities as the planetary rotation period and emission temperature but also provides estimates for such equally important but more elusive parameters as the static stability and vertical mixing coefficient.

Although the table is designed for nominal comprehension without external reference, this appendix also provides explanatory notes for the tabulations, including a brief summary of formulae for their simple application, and a complete reference listing of published data sources. Although the estimation of error bounds is an essential part of the observational assessment of planetary parameters, they have been omitted from the table not only for economy of space but also to avoid their misrepresentation out of context of the real uncertainties (e.g., in many cases where systematic errors associated with model dependent assumptions may exceed estimates of the formal statistical error). Some assessment of the precision of individual tabulations is

given in the explanatory notes. In general numbers have usually been rounded so that the claimed uncertainties affect at most the last decimal place reported. For critical applications, however, users are urged to consult the references to obtain a further account of assumptions and limitations as well as to confirm accurate citation!

#### ORBITAL PARAMETERS

The mean solar distance  $R$  (or semi-major axis) and the orbital eccentricity are specified as the (rounded) values for the osculating elements on 1987 January 5 as tabulated in *The Astronomical Almanac* (1986). (These describe the unperturbed two-body orbit that the planets would follow if the perturbations imposed by their neighbors were to cease instantaneously.) Distances are given in Astronomical Units (with 1 A.U. =  $1.49598 \times 10^8$  km). For Venus, Earth, Mars, and Jupiter these parameters are nearly invariable to within the given precision over decadal time scales within the current epoch, although they change by much larger amounts over several centuries (cf. Ward, 1974). For Saturn (and Titan), Uranus, and Neptune the osculating mean distances change by as much as a few tenths of a percent. Their eccentricities mostly change by no more than ten percent of their tabulated values with the notable exception of Neptune, whose osculating eccentricity has varied in recent years between 0.004 and 0.009. The numbers in the table have been rounded so that secular variation affects at most only the last decimal place reported.

The orbital periods  $\tau_{\text{orb}}$  (in days = 86,400 s and tropical years = 365.24 days), measured with respect to the fixed stars and rounded to five significant figures, are from Allen's *Astrophysical Quantities* (1973). For the outer planets, these are slightly shorter than their two-body Kepler period about the Sun, owing to the perturbing influence of their neighbors in inferior orbits.

The perihelion and Southern Summer Solstice dates (the first for each planet since 1985 May) and the  $L_S$  angle at perihelion are compiled or extrapolated from data in Allen (1973) and *The Astronomical Almanac*. ( $L_S$  is the planetocentric longitude of the sun measured eastward in the plane of the orbit from the ascending node on its equatorial plane, so that the Vernal Equinox corresponds to  $L_S = 0$  deg.) The Southern Summer Solstice corresponds to the time for which  $L_S = 270$  deg and was chosen for reference here because of its apparent relevance to the Martian global dust storms, the Voyager approach to Uranus encounter, and coincidentally, with possible arrival times planned for Galileo at Jupiter and Cassini at Saturn. (Projected calendar dates for perihelion and Solstice for the Jovian planets are reported here in tenths of years but may be in error by as much as 1% of their orbital periods.)  $L_S$  values at perihelion are given to facilitate the estimate of calendar dates for any  $L_S$  but may be in error by as much as a degree for the Jovian planets.

The obliquity is the inclination of a planet's equator to its orbital plane. The tabulated values are from *The Astronomical Almanac* (1986) and refer to the current epoch.

## PLANETARY ROTATION AND PHYSICAL STRUCTURE PARAMETERS

The sidereal rotation period  $\tau_{\text{rot}}$  is measured with respect to the fixed stars. The value for the Earth is from the *Astronomical Almanac* (1986). The rotation period for Mars has been derived from telescopic observations of the transit of surface features (cf. Ashbrook, 1953). The sidereal rotation of Venus has been determined from radar measurements of its surface (Shapiro et al., 1979). Jupiter's rotation is determined from measurements of its decametric radio emission (cf. Duncan, 1971). Titan's is taken to be the same as its orbital period about Saturn (as reported by Davies et al., 1980), assuming that its rotation is tidally locked to the planet. The rotation periods for Saturn (Desch and Kaiser, 1981) and Uranus (Warwick et al., 1986) are based on measurements of their periodic radio emissions by the Voyager Planetary Radio Astronomy investigation. The rotation period for Neptune can at present be only crudely estimated from determinations of its oblateness and gravitational moment according to principles briefly summarized below or from photometric observations of atmospheric periodicities (cf. Hubbard, 1986 and Belton et al., 1981). The periods for Venus and Uranus are given with negative signs to indicate that they rotate in a retrograde sense with respect to the pole that lies to the north of the invariable plane of the Solar System.

The equatorial radius values for Earth, Mars, Venus and Titan are referred to their solid surfaces and rounded to four significant figures. The value for the Earth is taken from *The Astronomical Almanac* (1986). The value for Mars is derived from a study by Christensen (1975) employing occultation, radar, spectral, and optical measurements. The Venus radius has been determined by Pettengill et al. (1980) using Pioneer Venus radar altimetry. The appended altitude for the haze level is derived from Pioneer Venus cloud photopolarimeter limb scan measurements by Lane and Opstbaum (1983). Titan's radius has been derived from Voyager radio occultation measurements, assuming that the satellite is spherical (Lindal et al., 1983). The indicated altitude of the main haze level on Titan corresponds to the elevation of its optical limb as measured by Voyager imaging (Smith et al., 1981). Tabulated values for the equatorial radius of the Jovian planets (again rounded to four significant figures) refer to the 1-bar pressure level of their atmospheres. Values for Jupiter and Saturn have been derived by Lindal et al. (1981, 1985) from a calculated geodetic fit to Voyager radio occultation measurements at different latitudes. The 1 bar equatorial radius value for Uranus has been derived by Hubbard (1984b) from stellar occultation observations of the planet by Elliot et al. (1981). The result agrees with Voyager imaging measurements (25,600 to 25,700 km, as reported by Smith et al., 1986) for the visible cloud deck which, according to Voyager radio science, is expected to reside at about 1.3 bar (Tyler, et al., 1986). The 1 bar radius for Neptune has been derived from stellar occultation data by Hubbard et al. (1985).

The oblateness  $\epsilon = (a_e - a_p) / a_e$  is a measure of the fractional difference between a planet's equatorial and polar radii. The tabulated value for Earth is from *The Astronomical Almanac* (1986), rounded to five significant figures. The Mars (optical) oblateness is from the study by Christensen (1975). Measurements of Venus altimetry by Pettengill et al. (1980) suggest that its oblateness is less

than about  $10^{-5}$ . The Titan oblateness is presently unknown. The geometrical oblateness of the fluid envelopes of the Jovian planets serves as an important measure of their rotational and gravitational structure. (Table sources are cited below together with a brief discussion of the inferred relationships to other parameters. Question marks follow the tabulated values for Saturn and Uranus as an indication of slight discrepancies between optical measurements and dynamical inference. The oblateness, once determined, provides a simple relationship between the planetocentric latitude coordinate  $\phi_c$ , measured along the oblate surface with respect to the center of the planet, and the planetographic coordinate  $\phi_g$ , measured with respect to the local normal to the same surface:

$$\tan \phi_g = (1-\epsilon)^{-2} \tan \phi_c \quad . \quad (1)$$

The differences between the two coordinates at midlatitudes on the Jovian planets are sufficiently great to warrant careful discrimination in reference to published results of atmospheric observations.

The gravitational parameter  $\underline{GM}$  is the product of the gravitational constant and planetary mass (in  $\text{km}^3 \text{s}^{-2}$ ). (Although  $G \cong 6.673 \times 10^{-3}$  is known to less than four-place precision, the product for most of the planets is now known to much greater accuracy.) Tabulated values for Earth, Mars, and Venus are taken from *The Astronomical Almanac* (1986). The value for Titan is from the tracking of Voyager radio science data reported by Tyler et al. (1981). Values for Jupiter (Null, 1976) and Saturn (Null et al., 1981) have been derived from the analysis of radio tracking data from the Pioneer spacecraft which, because of its close-encounter geometry with the two planets, provides the best available determination. The GM value for Uranus is a new result of the Voyager encounter (Tyler et al., 1986). The value for Neptune is that reported by Gill and Gault (1968) based on an analysis of the motion of Triton.

$\underline{J_2}$  and  $\underline{J_4}$  are the two lowest order coefficients in the multipole expansion expression for (an axially symmetric) planetary gravitational potential. Including the centrifugal potential associated with planetary rotation this expression may be written as

$$V(r, \phi) = -\frac{GM}{r} \left\{ 1 - \left[ \sum_{i=1}^{\infty} J_{2i} (a/r)^{2i} P_{2i}(\mu) \right] + (q/3)(r/a)^3 [1 - P_2(\mu)] \right\} \quad (2)$$

where  $(r, \phi)$  denote radial and (planetocentric) latitudinal coordinates,  $a$  is the normalizing radius for the expansion,  $\mu = \sin \phi$ ,  $q = \Omega^2 a^3 / GM$  ( $\Omega = 2\pi / \tau_{\text{rot}}$  is the planetary rotation frequency), and  $P_{2i}(\mu)$  denotes the  $(2i)$ th Legendre polynomial with  $P_2(\mu) = 1/2(3\mu^2 - 1)$ ,  $P_4(\mu) = (35/8)\mu^4 - (30/8)\mu^2 + 3/8$ , etc. Then to second order in the expansion:

$$\begin{aligned} V(r, \phi) = -\frac{GM}{r} \{ & 1 - J_2 (a_e/r)^2 [(3/2)\sin^2 \phi - 1/2] \\ & - J_4 (a_e/r)^4 [(3/8)\sin^4 \phi - (30/8)\sin^2 \phi + 3/8] \\ & + (q/2)(r/a_e)^3 \cos^2 \phi \} \end{aligned} \quad (3)$$

where the equatorial value  $a_e$  has been taken as the normalizing radius. Evaluating the potential at the equator ( $a_e, 0^\circ$ ) and the pole ( $a_p, 90^\circ$ ) and then equating the two results to solve for the relationship between  $a_e$  and  $a_p$  on an equipotential surface yields (for  $\epsilon \ll 1$ )

$$\epsilon \cong (3J_2/2 + q/2)(1 + 3J_2/2 - q/2) + 5J_4/8 \quad (4a)$$

or (to order  $J_2$ ):

$$\epsilon \cong 3J_2/2 + q/2 . \quad (4b)$$

Alternatively, the relationship for the rotation period in terms of  $\epsilon$ ,  $J_2$ , and  $J_4$  may be written as

$$\tau_{\text{rot}} = 2\pi \left[ \frac{a_e^3 (1-\epsilon)(1+3J_2/2)}{2GM(\epsilon - 3J_2/2 - 9J_2^2/4 - 5J_4/8)} \right]^{1/2} \quad (5a)$$

or (to lowest order in  $J_2$  and  $\epsilon$ ):

$$\tau_{\text{rot}} = 2\pi [a_e^3 / 2GM(\epsilon - 3J_2/2)]^{1/2} \quad (5b)$$

Clearly the nondimensional specification of the  $J_2$  and  $J_4$  coefficients as employed in Eq. (2) requires the adoption of a particular value for the normalizing radius  $a$  which, for various historical reasons, is often slightly different from atmospheric reference values for the equatorial radius. It has been traditional, for example, to employ a normalizing radius of 60,000 km for referencing the gravity moments of Saturn. For the purpose of specifying  $J_2$  and  $J_4$  for the Jovian planets in the present table, the published values have been renormalized to a reference radius equal to the tabulated 1 bar equatorial value. Gravitational studies of terrestrial planets often employ coefficient expansions with different normalizations, often including "off-diagonal" tesseral harmonics in addition to the zonal harmonics. Consequently, great care must be taken in comparing these for different planets (or different representations of a single planet.)

Tabulated  $J_2$  and  $J_4$  values for the Earth are from *The Astronomical Almanac*.  $J_2$  for Mars is taken from the analysis of combined tracking data for the Viking and Mariner 9 spacecraft by Gopcyński et al. (1977). (The Mars  $J_4$  value is omitted since it appears to be smaller than one of the second order tesseral harmonic coefficients.)  $J_2$  for Venus is taken from the analysis of tracking data for the Pioneer Venus orbiter by Ananda et al. (1980).  $J_2$  and  $J_4$  for Jupiter (Null, 1976) and Saturn (Null et al., 1981) are from the gravity analysis of the Pioneer 10 and 11 tracking data. The  $J_2$  and  $J_4$  values for Uranus have been derived by Elliot et al. (1981) from stellar occultation determinations of the precession of the planet's rings.  $J_2$  and  $J_4$  as determined in this way are inferred in proportion to the square root of GM. The table values reflect a renormalization of the results of Elliot et al. in terms of both the tabulated radius  $a_e$  and the Voyager determination of GM. The  $J_2$  value for Neptune is the (radius-renormalized) value derived by Harris (1984) from considerations of its spin-orbit coupling with Triton.

Accurate determinations of  $GM$ ,  $J_2$ , and  $J_4$  permit the dynamical inference of oblateness as indicated, for example, by Eq. (4). This relation assumes deformable fluid envelopes in hydrostatic balance. Solid planets may have non-equipotential surfaces with geometrical flattening different from the dynamical value inferred from Eq. (4), although in the case of the Earth these differences are small. For the Jovian planets, optical measurements of the geometrical oblateness provide an important check on the dynamical calculation. In the case of Jupiter the agreement is quite good. The tabulated number is the (rounded) value from a calculation by Lindal et al. (1981, 1985) which fits Voyager radio occultation data at various latitudes to the dynamical flattening of equipotential surfaces including the effects of the differential winds observed at cloud level. The same number also agrees with a stellar occultation measurement of Jupiter's oblateness by Hubbard (1977) and with the dynamical value inferred from Eq. (4), to within reported error limits. Determinations of Saturn's oblateness are more problematic. Equation (4), together with the tabulated values for  $GM$ ,  $J_2$ , and  $J_4$  yields  $\epsilon = 0.0963$ . An optical measurement from Pioneer 11 imaging photopolarimeter data yields the value  $0.088 \pm 0.006$  (Gehrels et al., 1980). Analysis of the geodetic fit to several radio occultation measurements by Lindal et al. (1985) yields  $0.09796 \pm 0.00018$  and implies that the centrifugal potential associated with Saturn's equatorial jet produces a 100 km bulge above the reference geoid. The number for the present table is taken as their value, rounded to four places. The tabulated value for Uranus is from a geometric determination with stratoscope II photographs by Franklin et al. (1980) and agrees within error bounds with stellar occultation measurements by Elliot et al. (1981), although both are larger by slightly more than the reported errors from the dynamical oblateness of Eq. (4) using the new Voyager rotation period. The tabulated oblateness for Neptune is taken from the stellar occultation measurements reported by Hubbard (1985, 1986) and is as yet uncontested by any independent measure of planetary spin rate and dynamical flattening.

Measured values of  $GM$  and  $J_2$  as applied to the multipole expansion of Eqs. (2) and (3) are also useful for the calculation of the gravitational acceleration  $g$  on an oblate equipotential surface of a rotating planet. This is given by the magnitude of the gradient of the total gravitational plus centrifugal potential normal to the surface, i.e.

$$g = [(\partial v / \partial r)^2 + (r^{-1} \partial v / \partial \phi)^2]^{1/2} \quad (6)$$

evaluated for the radial distance between the planetary center and its elliptical figure. For small values of the oblateness  $\epsilon$  the ellipse equation gives

$$r = a_e(1 - \epsilon \sin^2 \phi) \quad (7)$$

Neglect of the  $J_4$  term, substitution of equation (3) into (6), and evaluation for the radial distance given by (7) yields, to first order in the small parameters  $\epsilon$ ,  $J_2$ , and  $q$ :

$$g = GM/a_e^2 [1 + 3J_2/2 - q + (2\epsilon - 9J_2/2 + q) \sin^2 \phi] \quad (8)$$

Evaluation at the equator and the pole gives

$$g_E = GM/a_e^2 (1 + 3J_2/2) - \Omega^2 a_e \quad (9)$$

and 
$$g_P = GM/a_e^2 (1 + 2\epsilon - 3J_2) \quad (10)$$

It is then convenient to rewrite Eq. (8) in terms of these last results as

$$g = g_E + (g_P - g_E) \sin^2 \phi \quad (11)$$

An estimate of the global area-weighted mean value for the gravitational acceleration may be obtained by the integration of the product of this last expression for  $g$  with the cosine of the latitude with the result

$$\langle g \rangle = g_E + (g_P - g_E)/3 \quad (12)$$

As already mentioned, planets with solid surfaces may exhibit small departures from equipotential geoids. Nevertheless, Eqs. (9) to (12) are good first-order approximations for the estimation of rotation and oblateness corrections to their surface gravities and have been employed in the determination of the tabulated values for the area-weighted mean  $\langle g \rangle$ . The difference between the polar and equatorial values  $\delta g_E^P$  is also given and may be used together with Eqs. (11) and (12) to estimate the gravitational acceleration at any latitude. In the case of Jupiter and Saturn, the most elaborate geodetic study published incorporating the Voyager radio occultation soundings is that of Lindal et al. (1985) and for these two planets the tabulated mean value has been derived by application of Eq. (12) to their results for the equatorial and polar gravities. All tabulated values for the gravitational acceleration have been rounded to three significant figures. Applications requiring the accurate determination of  $g$  on the Jovian planets should consider the appendix to Lindal et al. (1985) outlining the iterative computation of higher order corrections than are contained in the simple formulae provided here.

#### PLANETARY HEAT FLOW PARAMETERS

Tabulated values of the internal heating, albedo, and effective emission temperature provide important characterizations of the radiative-convective state of planetary atmospheres. These quantities are related by the heat balance relation:

$$E = I + L \quad (13a)$$

where 
$$E = 4\pi a_e^2 \sigma T_e^4, \text{ the power emission,} \quad (13b)$$

$$I = \pi a_e^2 F_\odot (R/R_E)^{-2} (1-A), \text{ the power insolation,} \quad (13c)$$

$$L = 4\pi a_e^2 F, \text{ the planet's internal luminosity,} \quad (13d)$$

with  $\sigma = 5.67 \times 10^{-5} \text{ mW m}^{-2}\text{K}^{-4}$  designating the Stephen-Boltzmann constant,  $T_e$  the effective blackbody emission temperature,  $F_\odot = 1.37 \times 10^6 \text{ mW m}^{-2}$  the solar flux constant at 1 A.U. (Willson et al., 1980),  $(R/R_E)$  the (heliocentric) planetary distance in A.U.,  $A$  the Bond albedo, and  $F$  the planet's internal heat flux (as power per unit area). These relations (13a-d) neglect oblateness effects and complications related to the optical obscuration of planetary rings which are important for Saturn (cf. Hanel et al., 1983). For the case of a planet with an internal source, knowledge of both  $T_e$  and  $A$  permits the inference of the planet's self-luminosity or internal heat flux

$$F = \sigma T_e^4 - F_\odot (R/R_E)^{-2} (1-A)/4 \quad (14)$$

The internal heat is also usefully characterized in terms of the ratio of emitted to insolated (or absorbed) power as

$$E/I = 4\sigma T_e^4 (R/R_E)^2 / [F_\odot (1-A)] \quad (15)$$

Some researchers refer to the internal heating in terms of the fraction of solar input, sometimes denoted as

$$Q = (E/I - 1) = L/I \quad (16)$$

A convenient expression for the conversion of internal heating in these terms to the flux (as power per unit area) may be written as

$$F = (E/I)^{-1} (E/I - 1) \sigma T_e^4, \quad (17)$$

For the case of a planet with negligible internal heat source (14) reduces to

$$T_e \cong (279\text{K}) [(1-A)(R/R_E)^{-2}]^{1/4} \quad (18)$$

so that knowledge of either one of  $T_e$  and  $A$  together with  $(R/R_E)$  permits the inference of the other.

The tabulated value for the very small but still measureable internal heat flux  $F$  for the Earth is from Zharkov and Trubitsyn (1978). Mars, Venus, and Titan may also have very small internal heating but it cannot be measured remotely from spacecraft. The internal heating values for Jupiter (Hanel et al., 1981) and Saturn (Hanel et al., 1983) are from the analysis of Voyager IRIS measurements. The internal heat flux values for Uranus and Neptune have been estimated by Pollack et al. (1986) from a combination of ground based and Voyager data. The values for  $F$  in the present table have been obtained from their values of  $Q = (E/I-1)$  by application of equation (17). Tabulated values for  $E/I$  are from the same sources.

The Bond albedo  $A$  for the Earth is from a time and space mean analysis of observations from the Nimbus 7 spacecraft by Jacobowitz et al. (1984). The

Bond albedo for Mars is from an analysis of Mariner 9 infrared radiometric measurements by Kieffer et al. (1973) and confirms earlier ground-based photometric measurements by Irvine et al. (1968). The albedo value for Venus is from Pioneer Venus infrared radiometric measurements by Schofield and Taylor (1982). The value for Titan is derived from the effective temperature estimate of Lindal et al. (1983) by application of equation (18). Bond albedos for Jupiter and Saturn are from the Voyager IRIS analysis of Hanel et al. (1981, 1983). New estimates for Uranus and Neptune are from the work of Pollack et al. (1986).

The tabulated effective temperature values  $T_e$  are derived from the same sources as the Bond albedos, either by application of Eq. (18) (for Earth and Mars) or by reference to the separate specification of these by the authors of the papers cited for Titan, Jupiter, and Saturn. The tabulated value for the effective temperature of Neptune is taken as the upper limit estimated by Hanel et al. (1986) from Voyager IRIS measurements (rounded down to the nearest K) but is also within the error bounds on the number specified by Pollack et al. (1986).

It is also useful to evaluate the emission pressure level  $p_e$  corresponding to the emission temperature (sometimes called the "emission to space level") by reference to remotely retrieved or directly measured vertical structure profiles. The value for Earth is from the *U.S. Standard Atmosphere* (1976). The emission level for Mars is estimated in reference to an adopted model profile discussed below in the context of the surface temperature and pressure. The Venus emission level is determined by reference to *in situ* measurements of the pressure-temperature profile from Pioneer Venus probes by Seiff et al. (1980). The emission levels for Titan, Jupiter, Saturn, and Uranus are given by the same sources as referenced above for their measured albedo or effective temperature. The Neptune emission level is estimated from the radiative-convective model profile of Appleby (1986).

#### METEOROLOGICAL PARAMETERS

The 1-bar temperature  $T_{1b}$  is tabulated as a useful reference level for the vertical structure profile of the atmosphere. (For the Jovian planets this is nearly the deepest level which can be reliably retrieved from Voyager radio occultation or IRIS data and is therefore a useful benchmark for adiabatic extrapolation to lower levels.) For the Earth  $T_{1b}$  nearly coincides with its surface temperature (cf. *U.S. Standard Atmosphere*, 1976). (Of course the Martian atmosphere has no such level.) The 1-bar level for Venus is estimated from the in situ measurements by Seiff et al. (1980). The 1-bar retrievals for Titan (Lindal et al. 1983) and for Jupiter and Saturn (Lindal et al., 1981, 1985) are from Voyager radio occultation measurements. The 1-bar temperature for Uranus is from Voyager IRIS retrievals by Hanel et al., 1986. The Neptune value is estimated from radiative-convective models by Appleby (1986).

The surface temperature and pressure ( $T_s$  and  $p_s$ ) correspond to measurements of conditions at the solid surfaces of Earth, Mars, Venus, and Titan. The Earth values (from *U.S. Standard Atmosphere*) correspond to a time and space mean.

Determination of the time and space mean surface temperature for Mars is problematic. Diurnal, seasonal, as well as latitudinal variations are extreme, with temperatures ranging between 140 and 290 K. Kahn (1983) has assembled a cross section of diurnally averaged 20- $\mu\text{m}$  brightness temperatures over latitude and  $L_S$  using Viking IRTM measurements from Martin et al. (1979) together with otherwise unpublished data supplied by private communication. The area-weighted, time averaged temperature for this cross section is about 207 K. Although the 20- $\mu\text{m}$  channel is thought to provide a good measure of surface temperatures, this average may represent a slight underestimate of the actual mean surface value owing to the effect of measurements taken at non-zero emission angles. Although no Martian standard atmosphere is available, in a theoretical study of diurnal tides on Mars, Zurek (1976) offers a simple empirical model for the basic state temperature in the form

$$T_m(p) = 145\text{K} + (T_s - 145\text{K}) \exp[-\gamma \ln(p_s/p)] \quad (19)$$

where  $T_m$  is the mean (altitude-dependent) temperature,  $T_s$  and  $p_s$  are the surface temperature and pressure, and  $\gamma$  is a parameter related to the lapse rate. The high-altitude limit for this model gives a good fit to Viking lander descent data (cf. Seiff and Kirk, 1977). Zurek uses  $T_s = 220$  K (as do many authors) and suggests that  $\gamma = 0.64$  gives a good match to atmospheric lapse rates inferred from Mariner 9 IRIS measurements (Hanel et al., 1972). If, for average clear-air conditions, the Martian surface temperature is raised by a very weak greenhouse associated with the  $\text{CO}_2$  absorption of its thin atmosphere, then the application of the Eddington approximation would indicate that

$$T_s = T_e (1 + 3\tau/4)^{1/4} \quad (20)$$

where  $T_e$  is the effective emission temperature and  $\tau$  the optical depth.  $\tau = 0.1$  may be taken as a lower limit value for the surface under clear conditions (according to Leovy, 1979). Then, with  $T_e = 210\text{K}$  as derived from the radiometric albedo measurement, Eq. (20) yields  $T_s = 214$  K and coincidentally agrees with the average of the "canonical" value of 220 K and the IRTM result of 207K. On this (admittedly somewhat ad hoc) basis, the value of  $T_s \sim 214$  is adopted for tabulation, although it is probably uncertain by as much as 8K from the actual time and space mean. (It is possible that on an average basis the radiative screening of the surface by residual air-borne dust largely compensates for the very weak greenhouse warming and produces a shallow inversion layer). Mars surface pressures also vary substantially with the seasons (because of the sublimation and evaporation of the South polar cap) and with the topographic elevation. A mean surface pressure value of  $p_s = 0.007$  is estimated from Viking lander data (Ryan et al., 1978), adjusted for elevation with respect to the Mars geoid (cf. Seiff and Kirk, 1977). With these choices for the surface temperature and pressure, Zurek's (1976) model for the Martian pressure-temperature profile is modified to read

$$T_{\text{Mars}}(p) = 145\text{K} + (214-145)\text{K} (p/0.007\text{bar})^{0.64} \quad (21)$$

and has been used to derive the tabulated emission pressure corresponding to  $T_e$ .

The surface temperature and pressure for Venus are estimated from the Pioneer Venus probe measurements of Seiff et al. (1980).

The Jovian planets have no solid surfaces (except for relatively small rocky cores). For these the  $T_s$  and  $p_s$  are given instead as the estimated condensation level for water which, as a result of the associated latent heating and differentiation of mean molecular weight might act as a kind of (permeable) surface of strong buoyancy contrasts. Condensation levels are estimated by simple application of the integrated Clausius-Clapeyron equation which specifies that the saturation vapor pressure  $e_s$  changes with temperature according to

$$e_s = e_o \exp[(L_c m_v / R^* T_o) (1 - T_o / T)] \quad (22)$$

where  $e_o = 0.00611$  bar is the saturation vapor pressure of water at the triple-state temperature  $T_o = 273$  K,  $L_c$  is the latent heat of condensation,  $m_v$  the molecular weight of vapor and  $R^* = N_o k_B$  is the universal gas constant. (For water the factor  $L_c m_v / R^* T_o \cong 20$ .) The saturation vapor pressure can be expressed in terms of the molar mixing ratio of water  $f_{H_2O}$  using the partial pressure relation

$$e_s = (p - e_s) f_{H_2O} \cong p f_{H_2O} \quad (23)$$

Neglecting the effects of latent heat and differentiated molecular weight on the adiabat, the temperature is assumed to increase with depth as

$$T = T_{1b}(p/1\text{bar})^{R/c_p} \quad (24)$$

where  $R$  is the gas constant for dry atmosphere and  $c_p$  the specific heat at constant pressure. (Both quantities are discussed below.) Then using (23) and (24) to eliminate  $e_s$  and  $T$  in (22) gives

$$f_{H_2O} = (0.00611\text{bar}/p) \exp\{20[1 - (273\text{K}/T_{1b})(1\text{bar}/p)^{R/c_p}]\} \quad (25)$$

for the variation of the saturated mixing ratio of water with depth. Above the lower base of the cloud condensation level, the water mixing ratio will be depleted with altitude as indicated by this last result and possibly more by the action of dynamics and microphysical processes (cf. Rossow, 1978). At sufficiently deep levels below the condensation level the molar ratio of the vapor is expected to be well mixed and approximately constant with increasing depth. The condensation level itself is expected to occur where the saturated mixing ratio as a function of the local temperature and pressure equals the value for the deep atmosphere. Unfortunately, the  $H_2O$  abundance for the deep atmospheres of the Jovian planets is unknown. The analysis of Voyager IRIS and ground based data by Bjoraker et al. (1986) suggests that at the 5 bar level it is a factor of 100 below the solar composition value. Levels below 7 bars are inaccessible to remote observation, however, so that condensation of larger molar fractions at deeper levels cannot be ruled out. Table values for  $T_s$  and  $p_s$  on the Jovian planets correspond to the solution of Eq. (25) for a molar ratio equal to three times the solar abundance value (cf. Cameron, 1982) so that  $f_{H_2O} \cong 3.7 \times 10^{-3}$ . This represents a solar enrichment factor comparable to that observed for  $CH_4$  on Jupiter and Saturn. (cf. Gautier and Owen,

1983; Buriez and de Bergh, 1981.) It must be emphasized, however, that this is intended only as an illustrative example of the condensation parameters in the absence of any direct knowledge of the deep atmosphere.

The three most abundant major gases measured (or inferred) for each planetary atmosphere are listed along with their molar fractions. The measured ratios for the Earth's atmosphere are taken from Allen (1973). The major gas fractions for Mars are from an analysis of Viking lander data by Owen et al. (1977). The Venus gas fractions are those recommended by von Zahn et al. (1983) from a consideration of both Pioneer Venus and Venera spacecraft data. The approximate gas ratios for Titan have been inferred from a combination of Voyager IRIS and radio science data. The tabulated values are those suggested by Samuelson et al. (1981). The molar fractions for Jupiter are those derived from Voyager IRIS measurements by Conrath et al. (1984) as a revision of an earlier study by Gautier et al. (1981), also using the inferred CH<sub>4</sub>/H<sub>2</sub> ratio of Gautier and Owen (1983). The molar fractions for Saturn are also taken from Conrath et al., 1984, together with the CH<sub>4</sub>/H<sub>2</sub> ratio of Buriez and de Bergh (1981). The tabulated hydrogen and helium mole fractions for Uranus are the approximate results of a preliminary analysis of Voyager IRIS data by Hanel et al. (1986). The hydrogen-helium mole fractions of the Neptune atmosphere await precise measurement but are assumed to be roughly the same as the solar mixture (cf. Gautier and Owen, 1983).

The complex radiative, chemical, morphological, and microphysical properties of clouds in planetary atmospheres are still largely unknown. The present tabulation merely specifies the leading chemical constituents for the (upper level) clouds of each atmosphere. The probable three-component nature of Martian clouds and condensates is discussed by Pollack et al. (1977). The Venus clouds were identified as a highly concentrated solution of H<sub>2</sub>SO<sub>4</sub> by Sill (1972) and Young and Young (1973). The haze and clouds of Titan are thought to be a complex mixture of hydrocarbons (cf. Kunde et al., 1981). The Jovian planets are thought to have both NH<sub>3</sub> and H<sub>2</sub>O clouds, but the observations are still incomplete. (The current status of the relevant studies is reviewed by West, Strobel, and Tomasko, 1986.) Various metallic compounds such as MgH and SiH<sub>4</sub> may condense as clouds at pressure levels greater than 5000 bar (cf. Gierasch and Conrath, 1985) but are completely inaccessible to observation. It is likely that temperatures on Uranus and Neptune are cold enough to also effect the condensation of methane (cf. Atreya and Romani, 1985).

The gas constant  $R$  is given as the ratio of (the universal gas constant)  $R^* = N_0 k_B = 8.314 \times 10^7 \text{ g cm}^2 \text{ s}^{-2} \text{ mol}^{-1}$  (where  $N_0$  is Avagadro's number and  $k_B$  is the Boltzmann constant) to the mean molecular weight per mole of the atmospheric gas mixture. The mean molecular weights and resulting value for  $R$  have been computed from a molar-weighted average as indicated by the inventory of major gas constituents specified by the references cited above.

$c_p/R$  is the ratio of the molar specific heat at constant pressure to the gas constant. This is computed as

$$c_p/R = m \sum_i (f_i/m_i) (c_p/R)_i \quad (26)$$

where  $f_i$  and  $m_i$  are respectively the molar fraction and molecular weight of the  $i$ th component and  $m = R^*/R$  is the mean molecular weight of the total mixture. (The expression is derived from the assumption of an ideal gas mixture with a total specific heat equal to the molar weighted average of the specific heats of each of the components.) In the classical (high temperature) limit  $(c_p/R)_i = (2+n)/2$  where  $n$  is the total number of (translational, rotational, and vibrational) degrees of freedom of the molecules. Thus the ratio  $(c_p/R)_i = 5/2, 7/2, \text{ or } 9/2$  in the classical limit for the case of a monatomic, diatomic, or triatomic gas respectively and is in good agreement with actual observations of the relevant gases at room temperature. (For pure methane, the ratio is taken to be 4.23 according to data in the 1980 CRC Handbook of Chemistry and Physics.) For the cold upper tropospheres of the Jovian planets, where  $T \lesssim 300$  K, the molecular partition of internal energy and therefore the specific heat is significantly temperature dependent. The statistical ortho-para alignment of the hydrogenic protons also varies with temperature and adjusts to local equilibrium within a lag time that can be either as long as  $10^9$  s or much shorter, depending upon the presence of various catalyzing agents in the aerosols. Conrath and Gierasch (1984) have made a careful assessment of these effects in the context of the observations for Jupiter and Saturn. The size of the temperature-dependent effects on  $(c_p/R)$  for hydrogen is displayed in their Figure 9. (A similar plot for a Jovian hydrogen-helium mix is given by Conrath, 1986.) In view of the apparent variations,  $c_p/R = 3.3$  is tabulated for Jupiter as a compromise between the minimum value for equilibrium hydrogen which would prevail for temperatures near the one bar level and the larger value in the high-temperature limit obtained near the 7 bar level below. For Saturn, Uranus, and Neptune, equilibrium hydrogen at the colder temperatures of their tropopause levels will have a higher  $c_p/R$  ratio than the high temperature limit and an intermediate value of 3.6 is therefore adopted for tabulation.

The dry adiabatic lapse rate is computed as  $\Gamma = g/c_p$  from the tabulated values for the mean acceleration of gravity  $\langle g \rangle$ , the gas constant  $R$ , and the ratio  $c_p/R$ . The tabulated results are rounded to the nearest tenth of a Kelvin per kilometer. This will be a slight overestimate of the true dry adiabatic lapse rate in the deep atmospheres of Venus and Titan owing to non-ideal gas effects there (cf. Seiff et al., 1980 and Lindal et al., 1983). On the Jovian planets, strong variations of  $g$  with latitude as outlined above will result in corresponding changes in the adiabat. Furthermore, variations in the hydrogen ortho-para spin state as well as the variation of the specific heat of a given state with temperature, will result in substantial changes in the dry adiabatic lapse rate with altitude.

The static stability  $S = \Gamma + \partial T/\partial z$  is a measure of the buoyant restoring force acting on a parcel of atmosphere undergoing vertical displacements. The corresponding frequency of stable vertical oscillations is given by

$$N = (gS/T)^{1/2} \quad (27)$$

and is called the Brunt-Väisälä frequency. The static stability will in general vary with both latitude and elevation. Tabulated values refer to

estimated averages over selected altitudes. The tabulated value for the Earth is from a global annual mass-weighted mean between 200 and 1000 mb computed by Stone and Carlson (1979). The tabulated value for Mars is estimated as an average over two logarithmic pressure intervals (or scale heights) above the surface by application of the model atmosphere Eq. (23). The result is in good agreement with results obtained by Mariner 9 IRIS retrievals (Hanel et al., 1972), radio occultation measurements (Rasool and Stewart, 1971), and *in situ* Viking Lander descent data (Seiff and Kirk, 1977). (During dust storm conditions, the static stability may be reduced by around a factor of two.) The static stability for Venus is estimated for levels just below the cloud level (at around 45 km altitude) from the *in situ* probe measurements of Seiff et al. (1980). The value for Titan is estimated as the mean tropospheric stability indicated by the radio occultation measurements of Lindal et al. (1983). Tropospheric static stabilities for the Jovian atmospheres are exceedingly difficult to measure. Voyager radio occultation retrievals for Jupiter are nearly indistinguishable from the dry adiabat at levels below 1 bar. There is, however, an indirect inference of Jupiter's effective static stability based upon a mixing length theory for the transport of the planet's internal heat in the presence of ortho-para hydrogen conversion processes. Conrath and Gierasch (1984) have concluded that the Brunt frequency for vertical oscillations with frozen composition in an adiabatic equilibrium mean structure (expected to prevail below the 600 mb level) is constrained to approximately  $2 \times 10^{-3} \text{ s}^{-1}$  at 1 bar. The application of Eq. (27) to this result, together with tabulated values for  $g$  and  $T_{1b}$  implies a static stability of  $0.03 \text{ K km}^{-1}$ . The mixing length model also implies a rapid reduction in the stability with increasing depth. The tabulated value may therefore be regarded as an upper limit for levels below 1 bar in the absence of other phase change processes. Radio occultation measurements of lapse rates on Saturn between the 0.7 and 1.3 bar level (Lindal et al., 1985) imply a static stability of approximately  $0.05 \text{ K km}^{-1}$  when compared with a dry adiabat for dry normal hydrogen with a 3:1 ortho-para ratio. Since an equilibrium mixture will have a higher specific heat and therefore a lower adiabatic lapse rate, this result may also be regarded as an upper limit to the actual stability at that level.

The scale height  $H = RT/g$  where  $R$  is the gas constant,  $T$  the local temperature, and  $g$  the local gravity corresponds to an  $e$ -folding pressure depth of atmosphere. Values are computed for all planets at both the emission level and the surface (or estimated water condensation level on the Jovian planets) using the respective entries in the table.

The meridional thermal gradient (in Kelvins per 1000 km) is a useful scaling parameter for the analysis of the zonal momentum balance associated with large scale flows. The tabulated value for the Earth is estimated from the equator-to-pole drop at 500 mb as depicted in the Northern Hemisphere Winter cross section of Lorenz (1967). The value for Mars is estimated for the 7.6 km altitude (where the pressure is half the surface value) from the thermal cross section of Mariner 9 IRIS retrievals presented by Pollack et al. (1981). The meridional thermal gradient for Venus is estimated for the level of the main cloud deck (near 100 mb or 65 km altitude) from Pioneer Venus radio occultation data presented by Newman et al. (1984). The meridional gradient for Titan is

estimated from the Voyager IRIS brightness temperature analysis of Flasar et al. (1981) and refers to the 100 mb level. Meridional thermal gradients for Jupiter and Saturn have not been directly observed except above the cloud levels where the thermal wind analysis of Pirraglia et al. (1981) suggest a reduction of flow speeds with altitude. Nevertheless, estimates based on their measurements are adopted for tabulation and refer to changes over the horizontal scale of the jets.

The radiative time constant

$$\tau_{\text{Rad}} = (c_p \rho T) / (\sigma T_e^4 / H) = (pT) / \Gamma \sigma T_e^4 \quad (28)$$

is equivalent to the ratio of the thermal energy content of the atmosphere (per unit volume) to the radiative heating rate for one local scale height (per unit volume). Values are computed at both the  $p_e$  and  $p_s$  levels using the required information in the table.

The vertical eddy mixing coefficient  $\kappa$  (a.k.a. diffusion, viscosity, or exchange coefficient) is one of the most notorious parameters ever to be employed in the atmospheric sciences. It represents an attempt to parameterize the transport of conserved quantities by analogy to molecular dissipation and suffers from vexing uncertainties as to its size, spatial variation, and differences in application to heat, momentum, and trace constituents. Nevertheless, it finds essential application to such apparently different subjects as boundary layer theory and stratospheric chemistry. Fixing attention on purely vertical transport in the absence of any external forces, the idea is to represent the conservation of some quantity  $J$  as

$$\rho \partial J / \partial t = -\partial / \partial z [\rho (wJ - \kappa \partial J / \partial z)] \quad (29)$$

where  $\rho$  is the density,  $t$  and  $z$  are time and altitude coordinates,  $w$  is vertical velocity and the mixing coefficient

$$\kappa = \langle w'J' \rangle / (\partial \langle J \rangle / \partial z) \quad (30)$$

where  $w'$  and  $J'$  are the eddy fluctuations of vertical velocity and  $J$ . (The angle brackets denote a suitably defined average.) The scaling of these relations suggests that

$$\kappa \sim w'D \quad (31)$$

and

$$\tau_e \sim D^2 / \kappa \quad (32)$$

where  $D$  is the characteristic vertical scale of the transport (often the scale height) and  $\tau_e$  denotes the eddy "turn-over" time scale. One person's eddy mixing is another's up (and down) draft. While for most applications meteorologists attempt to minimize their reliance on  $\kappa$  by explicit account of  $w$ , aeronomers often seek to absorb all vertical transport into a single eddy diffusion coefficient which includes large-scale motions as well as small-scale turbulence. (An excellent review of this subject from the aeronomical viewpoint is given by Hunten, 1975.) Horizontal transports are also sometimes parameterized with horizontal exchange coefficients. These are often much

larger (for global scales) than the vertical coefficients but are even more problematic and will not be considered any further here. For specific applications it is important to distinguish between the diffusion of heat and momentum, since certain atmospheric eddy motions may transport one more efficiently than the other. This difference is sometimes expressed in terms of the Prandtl number  $P$ , defined as the ratio of the momentum diffusion coefficient to the heat diffusion coefficient. Several studies have shown, however, that for many atmospheric applications the Prandtl number is of order unity. For example, in the terrestrial boundary layer  $P=0.7$ , according to Sutton (1953). Assuming this is the case, the eddy mixing coefficient for planetary atmospheres may be estimated from a variety of recipes applicable to specific types of observations. A number of similarity relations are given by Priestley (1959). One especially important application for rapidly rotating planets with solid surfaces is the analysis of the Ekman wind spiral within the lower boundary layer (cf. Holton, 1979). This theory accounts for the observed turning of the wind vector with altitude by 45 deg between the surface and the geostrophic level aloft within a characteristic depth

$$D_E = \pi(2\kappa/f)^{1/2} \quad . \quad (33)$$

(Here  $f$  is the Coriolis parameter and is defined below.) Inference of this characteristic Ekman depth therefore yields a value for the strength of the eddy mixing. For applications to the very different context of Jovian atmospheres, useful estimates of eddy mixing may be made by application of the mixing length theory for the transport of heat in stellar interiors (cf. Clayton, 1983). This specifies that the mixing required to support the internal heat flux  $F$  is given as

$$\kappa = H(FR^2T/c_p)^{1/3} \quad (34)$$

where it has been assumed that the mixing length is given by the pressure scale height  $H$ . Equations (33) and (34) are only two different examples out of many other methods for determining the eddy mixing coefficient including the theory of tidal waves, the diagnostic analysis of heat and momentum balances for observed winds and temperatures, and solutions of diffusion models for the best match to observed chemical tracer abundances.

The tabulated vertical eddy mixing coefficient for the Earth is estimated from the application of Eq. (33) to observed Ekman layer depths at Jacksonville, Florida by Brown (1970). The result is one-half the value recommended by Hunten (1975) based on aeronautical considerations. Above the terrestrial tropopause Hunten suggests that the mixing coefficient drops rapidly to a minimum of  $2500 \text{ cm}^2\text{s}^{-1}$ , then increases gradually with height, and this number is also appended in the table for stratospheric applications. Leovy and Zurek (1979) have used the Ekman layer theory to fit diurnally averaged wind and pressure variations on Mars observed by Viking Lander 2 and infer an eddy viscosity of about  $10^5 \text{ cm}^2\text{s}^{-1}$ . French and Gierasch (1979) have applied a viscous boundary layer model to the Martian polar vortex and obtain a good match of calculated surface stress to observations of eolian wind streak features in the polar region with the choice of  $10^6 \text{ cm}^2\text{s}^{-1}$ . The tabulated

value of  $5 \times 10^5$  for the Mars eddy mixing coefficient is taken as a compromise between these two results and is the same as the value adopted by Kahn (1983). An estimated upper limit for the eddy mixing coefficient in the Venus atmosphere is derived from the scaling analysis for a meridional circulation model for the equatorial super-rotation by Gierasch (1975). This value is in good agreement with the inference of  $\kappa = 1.3 \times 10^5$  from measurements of the vertical haze distribution observed by Pioneer Venus photopolarimeter limb scans as derived by Lane and Opstbaum (1983). The upper limit on the eddy mixing in Titan's atmosphere is derived by Flasar et al. (1981) as a diagnostic analysis of meridional flow balances implied by Voyager IRIS observations. Eddy mixing coefficients for the Jovian tropospheres are derived from the application of the mixing length expression of Eq. (34) to tabulated values for  $T_s$ ,  $p_s$ , and the internal heat flux. Moist convection (cf. Gierasch, 1976) and ortho-para conversion processes (cf. Conrath and Gierasch, 1984) may act to reduce the strength of these large mixing coefficients by several orders of magnitude on scales smaller than the horizontal eddies associated with the zonal flow. Lewis and Fegley (1984) have argued, however, that vertical motions associated with the zonal winds may themselves produce vertical transports corresponding to eddy coefficients of nearly the same size as predicted by Eq. (34). It is important to understand that in such a case the "weather" produces the mixing and not the reverse. As for the Earth, the statically stable stratosphere overlying the emission level on the Jovian planets will be associated with a region of greatly reduced mixing compared with that of the deep atmosphere. Conrath and Pirraglia (1983) have argued that the reduction of the cloud-top winds with altitude inferred from the thermal wind shear may be understood in terms of a forced mean meridional circulation with eddy friction and radiative damping. Flasar (1986) has pointed out that the implied vertical damping scale suggests that the time scales for both dissipative processes is of comparable magnitude. Eddy mixing coefficients for the lower stratospheres of Jupiter and Saturn are therefore estimated by application of Eq. (32) with  $\tau_e = \tau_{rad}$  and  $D = H$  as evaluated at the emission level. The results are comparable to eddy diffusion coefficients employed by Strobel (1986) to describe the vertical distribution of photochemical constituents.

The Coriolis parameter  $f = 2\Omega \sin\phi$  (where  $\Omega = 2\pi/\tau_{rot}$ ) is the component of planetary vorticity normal to the local level surface (for latitude  $\phi$  in planetographic coordinates). Tabulated values are determined for 30 deg latitude.

The beta parameter  $df/d(a\phi) = (2\Omega/a)\cos\phi$  is the local planetary vorticity gradient. Tabulated values are determined for the equator.

The characteristic weather length  $L$  is used here to denote an estimate of the horizontal wavelength of meteorological features (pressure, temperature, and wind variations) divided by  $2\pi$ . This amounts to a measure of the reciprocal horizontal (dimensional) wavenumber and is useful for estimating the horizontal derivative of meteorological field variables in the scaling analysis of the equations of motion. For the Earth, the tabulated value for  $L=1000\text{km}$  is chosen as a characteristic measure of the scale of zonal midlatitude variations in temperature and pressure (high and low centers). It corresponds to a

midlatitude zonal wavenumber of 6 at the high altitude jet stream latitude (around 30 deg) as evident in hemispheric isobaric and isothermal cross-sections (e.g. Palmen and Newton, 1969). The same zonal wavenumber is evidenced in the spacing of midlatitude cloud forms as apparent, for example, in the southern hemisphere of the "blue marble" Apollo 8 photo of Earth from space. Thus  $L = a_e \cos(30^\circ)/6 \approx 920\text{km}$ . The length scale may also be estimated as the peak-to-peak separation of the northern and southern hemispheric 500 mb jet streams (both at around latitude 30 deg for solstice conditions, as depicted by Mintz, 1954), again divided by  $2\pi$ . By this meridional reckoning, therefore,  $L = 2(30^\circ/180^\circ)(a_e/2) = 1060\text{km}$ , in agreement with the zonal value. For Mars, L can be similarly estimated from the observed zonal wavenumber 4-6 associated with the passage of high and low pressure centers at the Viking Lander 2 site (Ryan et al., 1978). Then for Mars  $L = a_e \cos(48^\circ)/4 \approx 600\text{km}$ . A meridional estimate of the length scale on Mars may be inferred from the thermal wind field presented by Pollack et al. (1981). This shows a high-altitude jet stream at latitude 50 deg so that by analogy to the estimate for the Earth  $L = 2(50^\circ/180^\circ)(a_e/2) = 940\text{km}$  in fair agreement with the zonal determination. The tabulated length scale for Venus,  $L = 6000\text{km}$ , is inferred from the visually obvious zonal wavenumber 1 "Y-feature" in the clouds (Belton et al., 1976). Thus, the length scale for Venus is the same as the planet's radius. This is also consistent with the qualitative character of the zonal wind profile with latitude: a single super-rotating jet from pole-to-pole, symmetric about the equator. (Cloud tracked wind data presented by Rossow in 1985 also shows evidence for superimposed mid-latitude jets which may be associated with a secondary smaller length scale.) Voyager IRIS measurements of meridional thermal gradients on Titan are the only presently available evidence for atmospheric motions there, and show no sign of longitudinal variation. The analysis of these data by Flasar et al. (1981) suggest the presence of a cyclostrophic flow regime similar to that observed on Venus. This inference and the qualitatively monotonic equator-to-pole thermal gradient tentatively suggests a characteristic length scale for Titan equal to its radius, so that L is estimated to be  $\sim 3000\text{ km}$  but is sufficiently uncertain to warrant a question mark. The length scale for Jupiter is estimated as the width of a jet-stream pair (as measured, for example, by Limaye, 1986) divided by  $2\pi$ . The same estimation method is applied to Saturn, with observations reported by Ingersoll et al. (1984). Smith et al. (1986) have presented a latitudinal extrapolation of Voyager imaging measurements of drift speeds on Uranus suggesting a single prograde jet between 20 deg latitude and the pole. Taking this interval as a measure of one-half wavelength implies a horizontal scale  $L = 10,000\text{km}$  for Uranus. Horizontal scale measurements for Neptune must await the Voyager encounter in 1989.

The characteristic weather speed is given for both midlatitude and equatorial locations. (A plus sign designates prograde flow with respect to the planet's rotation, a minus sign retrograde flow.) Values for the Earth are estimated as the mean of Northern Winter and Southern Summer measurements at the 500mb level reported by Mintz (1954). The midlatitude Mars value is estimated from the thermal wind cross section of Pollack et al. (1981) as the average of the jet maxima at the 7.6 km (half pressure) altitude in the Northern and Southern Hemispheres. No equatorial wind measurement is available for Mars. Venus wind speeds are from cloud-tracked drift measurements reported by Rossow (1985).

These are independently confirmed by Doppler tracking measurements of the Pioneer Venus probes (Counselman et al., 1980) and the cyclostrophic wind analysis of radio occultation data by Newman et al. (1980). The Titan wind speed at 45 deg latitude and the 100mb (tropopause) level is taken from the thermal wind analysis of Flasar et al. (1981). Wind speeds for Jupiter (estimated from results by Limaye, 1986) and Saturn (from Ingersoll et al., 1984) are for the cloud-tracked wind level (probably no more than a scale height above the 1-bar level). The midlatitude wind speed for Uranus is taken as the maximum of the extrapolated fit to Voyager cloud tracked wind measurements given by Smith et al. (1986). No equatorial wind speeds for Uranus are available although the extrapolation of available data suggest retrograde velocities there. The tabulated wind speed for Neptune is from differential drift rates implied by atmospheric periodicities reported by Belton et al. (1981). This is assumed to apply to midlatitudes but is of uncertain interpretation.

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ASTRONOMICAL, PHYSICAL, AND METEOROLOGICAL PARAMETERS FOR PLANETARY ATMOSPHERES

	EARTH	MARS	VENUS	TITAN	JUPITER	SATURN	URANUS	NEPTUNE
Mean solar dist (AU)	1.0000	1.5237	0.7233	9.54	5.203	9.54	19.2	30.2
Orbital eccentricity	0.017	0.093	0.007		0.048	0.053	0.046	0.009
Orbital period $\tau_{orb}$	365.26d 1.0000yr	686.98d 1.8809yr	224.70d 0.61521yr		4332.6d 11.862yr	10759d 29.458yr	30685d 84.014yr	60189d 164.79yr
Perihelion Date	1986Jan2	1986Sep25	1985Oct6		1987.48	2003.5	2050.3	2042.0
$L_S$	281	251	253		57	278	187	357
So. Summer Solstice	1985Dec22	1986cOct25	1985cOct17	2002.6	1994.6	2002.6	1985.8	2001.8
Obliquity	23.45°	25.19°	177.3°	-	3.12°	26.73°	97.86°	29.56°
Rotation period $\tau_{rot}$	23 <sup>h</sup> 56 <sup>m</sup> 04 <sup>s</sup>	24 <sup>h</sup> 37 <sup>m</sup> 23 <sup>s</sup>	-243.01 <sup>d</sup>	15.945 <sup>d</sup>	9 <sup>h</sup> 55 <sup>m</sup> 30 <sup>s</sup>	10 <sup>h</sup> 39 <sup>m</sup> 24 <sup>s</sup>	-17 <sup>h</sup> 14 <sup>m</sup>	-15-18 <sup>h</sup>
Eqtr radius $a_e$ (km)	6378	3397	6051 <sup>[haze]</sup> <sub>@+84</sub>	2575 <sup>[haze]</sup> <sub>@+240</sub>	71,490	60,270	25,660	24,820
Oblateness $\epsilon=1-a_p/a_e$	0.00335	0.0059	$\leq 0.00001$	-	0.0649	0.0980?	0.022?	0.022
GM ( $\text{km}^3\text{s}^{-2}$ )	3.986x10 <sup>5</sup>	4.283x10 <sup>4</sup>	3.249x10 <sup>5</sup>	8.976x10 <sup>3</sup>	1.267x10 <sup>8</sup>	3.794x10 <sup>7</sup>	5.794x10 <sup>6</sup>	6.88x10 <sup>6</sup>
$J_2 \times 10^6$ @ $a_e$	1082.6	1956	6	-	14694	16330	3490	4400
$J_4 \times 10^6$	-1.6	-	-	-	-580	-920	-30	-
Mean $\langle g \rangle  \delta g_E^p $ ( $\text{cm s}^{-2}$ )	980 5	371 3	887 0	135 0	2440 389	1000 318	875 51	1090 21
Internal heat flux $F$ ( $\text{erg cm}^{-2}\text{s}^{-1}$ )	62	-	-	-	5400	2000	<150	~450
Emission/Insolation	1.0002	1.000	1.000	1.000	1.7	1.8	<1.3	-2.8
Bond albedo	0.306	0.25	0.76	0.22	0.34	0.34	0.34	-0.28
Effective emission temperature $T_e$ (K)	255	210	229	85	124	95	59	59
pressure $p_e$ (bar)	0.5	0.006	0.02	1	0.4	0.3	0.4	0.5
1-bar temp (K)	288	-	346	86	165	134	78	75?
Surface [or Jovian H <sub>2</sub> O] cld if 3@mix								
temperature $T_S$ (K)	288	-214	731	94	[294]	[313]	[366]	[360]
pressure $p_S$ (bar)	1.013	0.007	92	1.5	[7]	[21]	[260]	[283]
Major gases [molar ratios]	N <sub>2</sub> [0.781] O <sub>2</sub> [0.209] Ar[0.009]	CO <sub>2</sub> [0.953] N <sub>2</sub> [0.027] Ar[0.016]	CO <sub>2</sub> [0.965] N <sub>2</sub> [0.035] Ar[70ppm]	N <sub>2</sub> [0.822?] Ar[0.166?] CH <sub>4</sub> [0.060?]	H <sub>2</sub> [0.899] He[0.099] CH <sub>4</sub> [0.002]	H <sub>2</sub> [0.963] He[0.033] CH <sub>4</sub> [0.004]	H <sub>2</sub> [~0.85] He[~0.15]	H <sub>2</sub> [~0.9?] He[~0.1?]
Clouds	H <sub>2</sub> O	Dust, H <sub>2</sub> O, CO <sub>2</sub>	H <sub>2</sub> SO <sub>4</sub>	hydrocarbons	NH <sub>3</sub> , H <sub>2</sub> O?	NH <sub>3</sub> , H <sub>2</sub> O?	CH <sub>4</sub> , NH <sub>3</sub> , H <sub>2</sub> O?	CH <sub>4</sub> , NH <sub>3</sub> , H <sub>2</sub> O?
Gas const $R$ ( $\text{cm}^2\text{s}^{-2}\text{K}^{-1}$ )	2.87x10 <sup>6</sup>	1.92x10 <sup>6</sup>	1.91x10 <sup>6</sup>	2.9x10 <sup>6</sup>	3.71x10 <sup>7</sup>	3.89x10 <sup>7</sup>	3.6x10 <sup>7</sup>	3.8x10 <sup>7</sup>
$c_p/R$	3.5	4.4	4.5	3.6	3.3	3.6	3.6	3.6
Dry adiabatic lapse rate $\Gamma=g/c_p$ ( $\text{K km}^{-1}$ )	9.8	4.4	10.5	1.3	varies with altitude and ortho-para H <sub>2</sub> ratio			
Static stability $S=\Gamma+\partial T/\partial z$ ( $\text{K km}^{-1}$ )	4.6	?	?	0.7	<0.0?	$\leq 0.05$	?	?
		var w dust	near clds		below 1-bar level			
Scale height $H$ (km)								
@ emission level $p_c$	7.5	11	4.9	18	19	37	24	21
@ surface [or H <sub>2</sub> O] $p_s$	8.4	11	16	20	45	122	150	124
Meridional thermal gradient ( $\text{K}/10^3\text{km}$ )	-4	-8	-6	-1	<±2?	<±3?	?	?
Radiative time const (s)								
$\tau_{Rad}=pT/\Gamma\sigma T_e^4$ @ $p_e$	5x10 <sup>6</sup>	3x10 <sup>5</sup>	3x10 <sup>5</sup>	2x10 <sup>9</sup>	2x10 <sup>8</sup>	9x10 <sup>8</sup>	5x10 <sup>9</sup>	5x10 <sup>9</sup>
@ $p_s$	1x10 <sup>7</sup>	3x10 <sup>5</sup>	4x10 <sup>9</sup>	3x10 <sup>9</sup>	8x10 <sup>9</sup>	2x10 <sup>11</sup>	2x10 <sup>13</sup>	2x10 <sup>13</sup>
Vertical eddy mixing coeff $\kappa_V$ ( $\text{cm}^2\text{s}^{-1}$ )	5x10 <sup>4</sup> trop >2500 aloft	5x10 <sup>5</sup>	$\leq 2 \times 10^5$	$\leq 10^3$	6x10 <sup>8</sup> deep 2x10 <sup>4</sup> aloft	8x10 <sup>8</sup> deep 2x10 <sup>4</sup> aloft	<2x10 <sup>8</sup>	2x10 <sup>8</sup>
Coriolis parameter $f(30^\circ)=2\pi/\tau_{rot}$ ( $\text{s}^{-1}$ )	7.292x10 <sup>-5</sup>	7.088x10 <sup>-5</sup>	-2.993x10 <sup>-7</sup>	4.561x10 <sup>-6</sup>	1.758x10 <sup>-4</sup>	1.638x10 <sup>-4</sup>	-1.012x10 <sup>-4</sup>	-1x10 <sup>-4</sup> ?
Beta parameter $\beta_e=2f/a_e$ ( $\text{cm}^{-1}\text{s}^{-1}$ )	2.29x10 <sup>-13</sup>	4.17x10 <sup>-13</sup>	9.89x10 <sup>-16</sup>	3.54x10 <sup>-14</sup>	4.92x10 <sup>-14</sup>	5.43x10 <sup>-14</sup>	7.88x10 <sup>-14</sup>	~9x10 <sup>-14</sup> ?
Characteristic weather length $L$ (km) (wavelength/2 $\pi$ )	1000	600	6000	3000?	2000	3000	10,000	?
speed $U$ @ mid-lat ( $\text{m s}^{-1}$ ) @ eqtr	+15 -4	+30 -	+90 +100	+40 +?	±50 +100	+100 +500	+200 -?	100? -