4.2.3.2 Planetary interiors

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4.2.3.2.1 Introduction

4.2.3.2.1.1 Symbols used

Table 1. List of symbols used.

Symbol	Definition	Unit
a	(Sub-planetary) equatorial radius	km
A	Principal equatorial moment of interia	kg m^2
b	Along-orbit equatorial radius	km
B	Principal equatorial moment of interia	kg m^2
c	Polar radius	km
c_p	Specific heat	J K^{-1} kg^{-1}
C	Principal polar moment of inertia	kg m^2
C_m	Mantle moment of inertia with respect to rotational axis	kg m^2
$C_{n,m}$, $S_{n,m}$	Gravitational coefficients of degree n and order m	-
g	Gravity	m s^{-2}
G	Gravitational constant	m^3 kg^{-1} s^{-2}
G_μ	Shear modulus	Pa
GM	Planetocentric constant	km^3 s^{-2}
I	Mean moment of inertia	kg m^2
J_n	Zonal gravitational coefficient	-
k	Thermal conductivity	W m^{-1} K^{-1}
k_2	Tidal potential Love number	-
k_f	Fluid potential Love number	-
K_S	Adiabatic bulk modulus	Pa
K_T	Isothermal bulk modulus	Pa
K_0T	Isothermal bulk modulus at reference state	Pa
K_0'	First pressure derivative of isothermal bulk modulus at reference state	-
l	Logarithmic volume derivative of Grüneisen parameter	-
m	Mass	kg
m_Fe	Iron mass	kg
M	Total mass	kg
M_c	Core mass	kg
p	Hydrostatic pressure	Pa
p_0	Reference pressure	Pa
p_base	Crust-mantle-transition pressure	Pa
p_c	Central pressure	Pa
p_cmb	Core-mantle-boundary pressure	Pa
p_th	Thermal pressure	Pa
q	Heat flux	W m^{-2}
Q_p	Tidal quality factor	-

Table 1. (continued) List of symbols used.

Symbol	Definition	Unit
r	Radial distance	m
R	Mean surface radius	m
R_{C}	Core radius	m
T	Temperature	K
T_0	Reference temperature	K
V_{P}	Longitudinal P-wave velocity	m s^{-1}
V_{S}	Transversal S-wave velocity	m s^{-1}
V_0	Molar volume	cm^3 mol^{-1}
x_{Fe}	Concentration of iron per unit mass	-
α	Thermal expansivity	K^{-1}
α_0	Thermal expansivity at reference state	K^{-1}
γ	Thermodynamic Grüneisen parameter	-
γ_0	Thermodynamic Grüneisen parameter at reference state	-
δ	Apparent tidal phase lag	deg
ϵ	Specific heat production rate	W kg^{-1}
χ_{Fe}	Mole fraction of iron $\in$ [0,1]	-
χ_{FeS}	Mass fraction of iron sulfide $\in$ [0,1]	-
χ_{S}	Mass fraction of sulfur $\in$ [0,0.365]	-
$\bar{\mu}$	Mean atomic mass	g $atom^{-1}$
Φ	Seismic parameter	m^2 s^{-2}
ρ	Density	kg m^{-3}
$\bar{\rho}$	Mean density	kg m^{-3}
ρ_0	Density at reference state	kg m^{-3}
ρ_{c}	Core density	kg m^{-3}
ρ_{m}	Mantle density	kg m^{-3}
ρ_{s}	Crust density	kg m^{-3}
ϱ_0	Specific mass of mineral at reference state	g cm^3
τ	Superadiabatic temperature gradient	K m^{-1}
θ	Moment of inertia	kg m^2
θ_{D}	Debye temperature	K
ω	Mean angular frequency of rotational and orbital period	s^{-1}

4.2.3.2.1.2 Overview

The internal structure and bulk composition of solar system bodies are key to understanding the origin and early evolution of the solar system. In general, solar system bodies are composed of rocks, metals, ices, fluids, and gases. While rock and metal are the main constituents of terrestrial, i.e. Earth-like, bodies in the inner solar system, more volatile components in form of ices and gases dominate in the outer solar system. The principal structure of terrestrial planet interiors can be described in terms of three-layer models, representing the metallic core, the silicate mantle and the crust. Since seismological observations of solar system bodies are only available for the Earth and, to a much lesser extent, the Moon, the average density and axial moment of inertia of a planet constrain the radial distribution of mass and subdivision into core, mantle, and crust layers. Additionally, large-scale topography and gravity data can be used to infer a planet's three-dimensional internal mass distribution, since the shapes of the physical surface and the external gravitational field are related to the radial density stratification and compositionally and/or thermally induced lateral density heterogeneities. Interior structure models aim at calculating (1) the volumes and masses of major chemical reservoirs that contribute to the bulk composition; (2) the depths to chemical and rheological discontinuities and mineral phase boundaries; and (3) depth variations of pressure, temperature, density, and composition. Since there are usually fewer constraints than unknowns, even basic interior structure models that would involve only two or three chemically homogeneous layers of constant density suffer from inherent non-uniqueness.

4.2.3.2.2 Basic equations and models

4.2.3.2.2.1 Two-layer structural models

The mean density of a two-layer spherical body is given by

$$\bar{\rho}R^3 = (\rho_\mathrm{c} - \rho_\mathrm{m})R_\mathrm{c}^3 + \rho_\mathrm{m}R^3 \tag{1}$$

from which the relative core radius R_c/R, core mass fraction M_c/M, and dimensionless mean moment-of-inertia factor (MoI) are obtained according to

$$\frac{R_\mathrm{c}}{R} = \left(\frac{\bar{\rho} - \rho_\mathrm{m}}{\rho_\mathrm{c} - \rho_\mathrm{m}}\right)^{1/3}, \tag{2}$$

$$\frac{M_\mathrm{c}}{M} = \frac{\rho_\mathrm{c}}{\bar{\rho}}\left(\frac{R_\mathrm{c}}{R}\right)^3 = \frac{\rho_\mathrm{c}}{\bar{\rho}}\frac{(\bar{\rho} - \rho_\mathrm{m})}{(\rho_\mathrm{c} - \rho_\mathrm{m})}, \tag{3}$$

and

$$\mathrm{MoI} = \frac{I}{MR^2} = \frac{2}{5}\left(\frac{(\bar{\rho} - \rho_\mathrm{m})^{5/3}}{\bar{\rho}(\rho_\mathrm{c} - \rho_\mathrm{m})^{2/3}} + \frac{\rho_\mathrm{m}}{\bar{\rho}}\right), \tag{4}$$

respectively, where the core radius R_c, the mantle density ρ_m, and the core density ρ_c are unknown.

For a rapidly rotating planetary body in hydrostatic equilibrium, the mean MoI as a measure for mass concentration toward the center can be obtained from the dimensionless polar moment-of-inertia factor, $C/M/R^2$, and the degree-two coefficient of the spherical harmonic representation of the gravitational field, J_2, by

$$\frac{I}{MR^2} = \frac{C}{MR^2} - \frac{2}{3}J_2 \tag{5}$$

where

$$J_2 = \frac{1}{MR^2}\left[C - \frac{A+B}{2}\right] = -C_{2,0} \tag{6}$$

is the gravitational oblateness and $A < B < C$ are the planet's principal equatorial and polar moments of inertia, respectively.

The gravitational acceleration g_r and hydrostatic pressure p_r as a function of radial distance from the center of the planet r are given by

$$\begin{aligned} g_r &= \frac{4}{3}\pi G r \rho_\mathrm{c} \\ &\quad 0 \le r \le R_\mathrm{c} \\ &= \frac{4}{3}\pi G r\left[\rho_\mathrm{m} + (\rho_\mathrm{c} - \rho_\mathrm{m})\left(\frac{R_\mathrm{c}}{r}\right)^3\right] \\ &\quad R_\mathrm{c} \le r \le R \end{aligned} \tag{7}$$

and

$$
\begin{aligned}
p_r &= \frac{4}{3}\pi G \rho_{\mathrm{m}} R_{\mathrm{c}}^3 (\rho_{\mathrm{c}} - \rho_{\mathrm{m}}) \left[\frac{1}{r} - \frac{1}{R}\right] + \frac{2}{3}\pi G \rho_{\mathrm{m}}^2 (R^2 - r^2) \\
&\quad R_{\mathrm{c}} \le r \le R \\
&= \frac{2}{3}\pi G \rho_{\mathrm{c}}^2 (R_{\mathrm{c}}^2 - r^2) + \frac{2}{3}\pi G \rho_{\mathrm{m}}^2 (R^2 - R_{\mathrm{c}}^2) + \frac{4}{3}\pi G \rho_{\mathrm{m}} R_{\mathrm{c}}^3 (\rho_{\mathrm{c}} - \rho_{\mathrm{m}}) \left[\frac{1}{R_{\mathrm{c}}} - \frac{1}{R}\right] \\
&\quad 0 \le r \le R_{\mathrm{c}}
\end{aligned} \tag{8}
$$

respectively, where G is the gravitational constant [02Tur].

4.2.3.2.2.2 Three-layer structural models

The structural equations of a three-layer spherical body, representing core, mantle, and crust, are given by

$$\bar{\rho} = \rho_{\mathrm{s}} + (\rho_{\mathrm{c}} - \rho_{\mathrm{m}}) \left(\frac{R_{\mathrm{c}}}{R}\right)^3 + (\rho_{\mathrm{m}} - \rho_{\mathrm{s}}) \left(\frac{R_{\mathrm{m}}}{R}\right)^3 \tag{9}$$

$$\frac{I}{MR^2} = \frac{2}{5}\left(\frac{\rho_{\mathrm{s}}}{\bar{\rho}} + \frac{\rho_{\mathrm{c}} - \rho_{\mathrm{m}}}{\bar{\rho}} \left(\frac{R_{\mathrm{c}}}{R}\right)^5 + \frac{\rho_{\mathrm{m}} - \rho_{\mathrm{s}}}{\bar{\rho}} \left(\frac{R_{\mathrm{m}}}{R}\right)^5\right) \tag{10}$$

where the core radius R_{c}, the crust-mantle radius R_{m}, the crust density ρ_{s}, the mantle density ρ_{m}, and the core density ρ_{c} are unknown.

4.2.3.2.2.3 Depth-dependent structural models

The construction of depth-dependent models of the interior structure relies on the assumption of a spherically symmetric planet in perfect mechanical and thermal equilibrium. Spherical models of the internal density distribution of planetary bodies are required to satisfy two constraints, the mean density $\bar{\rho}$ as derived from the total radius R and mass M and the mean moment of inertia I that can be inferred from the quadrupole moments of the gravitational field and the rotational dynamics of the spin axis. The following set of differential equations for mass m, iron mass m_{Fe}, mean moment of inertia θ, acceleration of gravity g, pressure p, and heat flux q can be derived from fundamental principles [07Soh]:

$$\frac{\mathrm{d}m}{\mathrm{d}r} = 4\pi r^2 \rho_r \tag{11}$$

$$\frac{\mathrm{d}m_{\mathrm{Fe}}}{\mathrm{d}r} = x_{\mathrm{Fe},r} \frac{\mathrm{d}m}{\mathrm{d}r} \tag{12}$$

$$\frac{\mathrm{d}\theta}{\mathrm{d}r} = \frac{8}{3}\pi r^4 \rho_r \tag{13}$$

$$\frac{\mathrm{d}g}{\mathrm{d}r} = 4\pi \mathrm{G} \rho_r - 2\frac{g_r}{r} \tag{14}$$

$$\frac{\mathrm{d}p}{\mathrm{d}r} = -\rho_r g_r \tag{15}$$

$$\frac{\mathrm{d}q}{\mathrm{d}r} = \rho_r \epsilon_r - 2\frac{q_r}{r} \tag{16}$$

where r is the radial distance from the center of the planet, G is the gravitational constant, ρ is the density, x_{Fe} is the concentration of iron per unit mass, and ϵ is the specific heat production rate. The subscript r indicates quantities that are local functions of pressure, temperature, and composition.

Heat is primarily carried by conduction across the stagnant outer portion of a planet's silicate shell and the top and bottom thermal boundary layers of mantle convection. The corresponding radial temperature gradient is given by

$$\frac{\mathrm{d}T}{\mathrm{d}r} = -\frac{q}{k_r} \tag{17}$$

where k is the thermal conductivity. Within the convective portion of the silicate shell and the outer liquid core shell, the temperature gradient can be approximated by the adiabatic temperature gradient [77Sta]

$$\frac{\mathrm{d}T}{\mathrm{d}r} = T\frac{\gamma_r}{K_{S,r}}\frac{\mathrm{d}p}{\mathrm{d}r} = T\frac{\gamma_r \rho_r}{\Phi_r}\frac{\mathrm{d}p}{\mathrm{d}r} = \left(\frac{\mathrm{d}T}{\mathrm{d}r}\right)_{\mathrm{ad}} \tag{18}$$

where $\gamma = \alpha K_S/\rho c_p$ is the thermodynamic Grüneisen parameter, α is the thermal expansivity, c_p is the specific heat, and K_S is the *adiabatic* bulk modulus defined as

$$\frac{1}{K_S} = \frac{1}{\rho}\left(\frac{\partial \rho}{\partial p}\right)_S \tag{19}$$

Finally, ρ, K_S, and the shear modulus G_μ are related to the seismic velocities

$$V_{P,r} = \sqrt{\frac{K_{S,r} + \frac{4}{3}G_{\mu,r}}{\rho_r}} \tag{20}$$

and

$$V_{S,r} = \sqrt{\frac{G_{\mu,r}}{\rho_r}} \tag{21}$$

of longitudinal P- and transversal S-waves, respectively, from which the seismic parameter

$$\Phi_r = \frac{K_{S,r}}{\rho_r} = V_{P,r}^2 - \frac{4}{3}V_{S,r}^2 \tag{22}$$

can be derived.

The set of basic differential equations (11) – (18) can be separated into two subsets that are coupled through the density ρ. The mechanical properties of the interior are calculated from Eqs. (11) – (15), while Eqs. (16) – (18) give the thermal structure of the model. These equations have to be supplemented with an appropriate equation of state to include pressure-induced compression and thermal expansion effects on the density and are discussed in more detail further below.

4.2.3.2.2.4 Boundary conditions

The set of basic differential equations (11) – (18) can be solved by numerical integration with respect to the following boundary conditions. The central boundary conditions at $r = 0$ are

$$m = 0;\ m_{\mathrm{Fe}} = 0 \quad ;\ \theta = 0 \quad ;\ g = 0 \quad ;\ p = p_{\mathrm{c}} \quad ;\ q = 0 \quad ;\ T = T_{\mathrm{c}} \tag{23}$$

The surface boundary conditions at $r = R$ are

$$m = M \quad ;\ m_{\mathrm{Fe}} = M_{\mathrm{Fe}} \quad ;\ \theta = I \quad ;\ g = g_{\mathrm{p}} \quad ;\ p = p_{\mathrm{p}} \quad ;\ q = q_{\mathrm{p}} \quad ;\ T = T_{\mathrm{p}} \tag{24}$$

While the mass M and the mean surface values of gravity g_{p}, pressure p_{p}, and temperature T_{p} have been derived from spacecraft and Earth-based observations, the surface heat flux q_{p} of most terrestrial bodies other than the Earth are unknown at present. Since there are three observational constraints on the model, the mass M, the radius R, and the mean moment of inertia I, respectively, three parameters are adjustable, the values of which are iteratively adjusted such that the observational constraints can be satisfied. These parameters are the central pressure p_{c}, the central temperature T_{c}, and the pressure at the core-mantle boundary p_{cmb}.

4.2.3.2.2.5 Equation of state

A convenient method to calculate density as a function of depth, first applied to seismological data from the Earth's deep interior, relies on the assumption of hydrostatic pressure (Eq. (15)) and adiabatic temperature (Eq. (18)) conditions and is sometimes referred to as the *Williamson-Adams* method honoring its authors. The resultant adiabatic density gradient

$$\frac{\mathrm{d}\rho}{\mathrm{d}r} = -\frac{\rho_r g_r}{\Phi_r} = {\Phi_r}^{-1}\frac{\mathrm{d}p}{\mathrm{d}r} \tag{25}$$

can be readily obtained from insertion of Eqs. (19) and (22) in (15) and rearranging terms. Later, a correction term was added to address temperature deviations from an adiabatic reference state. These are caused by thermal boundary layers of mantle convection and/or compositional changes and pressure-induced mineral phase transformations in a silicate mantle, giving rise to superadiabatic temperature gradients

$$\tau = \frac{\mathrm{d}T}{\mathrm{d}r} - \left(\frac{\mathrm{d}T}{\mathrm{d}r}\right)_{\mathrm{ad}} \tag{26}$$

that may profoundly affect the density stratification. The overall density gradient is then given by

$$\frac{\mathrm{d}\rho}{\mathrm{d}r} = {\Phi_r}^{-1}\frac{\mathrm{d}p}{\mathrm{d}r} + \alpha_r\rho_r\tau_r \tag{27}$$

and can be derived from the ambient density ρ and the seismic parameter Φ if adiabatic compression dominates (first term). In the event of perfect adiabaticity, τ will equal zero, and the superadiabatic correction (second term) would vanish.

To interpret density profiles inferred from seismological observations in terms of compositional changes and temperature anomalies, it is necessary to consider decompressed values of density. Reliable extrapolation to zero-pressure and room-temperature conditions require laboratory measurements of thermodynamic material properties experimentally determined at high pressures and temperatures. Extrapolation between different thermodynamic states represents a principal source of uncertainty in density modeling because of incomplete knowledge of the material parameters involved. For technical reasons, laboratory measurements are usually carried through under isothermal conditions.

The *isothermal* bulk modulus is then defined as

$$\frac{1}{K_T} = \frac{1}{\rho}\left(\frac{\partial\rho}{\partial p}\right)_T \tag{28}$$

and can be related to K_S by using the isothermal-adiabatic transformation

$$\frac{K_S}{K_T} = 1 + \gamma\alpha T \tag{29}$$

if the Grüneisen parameter γ, taken as $\gamma_0(\rho/\rho_0)^{-l}$ and $l = \text{const.}$ and thermal expansivity α are specified. A linearly pressure-dependent expression for K_T is frequently used, i.e.,

$$K_T = K_{0T} + \frac{\mathrm{d}K_{0T}}{\mathrm{d}p}p = K_{0T} + K_0'p \tag{30}$$

and index 0 referring to an arbitrarily chosen reference state at fixed (standard) pressure and temperature conditions p_0 and T_0, and constant first pressure derivative K_0'. Insertion of Eq. (30) in (28) and subsequent integration yields

$$p = \frac{K_{0T}}{K_0'}\left[\left(\frac{\rho}{\rho_0}\right)^{K_0'} - 1\right] \tag{31}$$

which is occasionally referred to as *Murnaghan's* equation. The corresponding density relative to its reference value is then given by

$$\frac{\rho}{\rho_0} = \left(1 + \frac{K_0'}{K_{0T}} p\right)^{1/K_0'} \tag{32}$$

The density distribution within a chemically homogeneous layer was extended by [52Bir] who suggested to employ a higher-order isothermal finite-strain equation of state to correct for pressure-induced compression and to apply temperature corrections through calculation of the thermal pressure contribution [81Sta].

For example, an isothermal *Birch-Murnaghan* equation truncated at third-order in strain involves K_{0T} and K_0' to correct for pressure-induced compression and the product $\alpha_0 \times K_{0T}$ to correct for thermal pressure effects, according to [80And]

$$p = \frac{3K_{0T}}{2}\left[\left(\frac{\rho}{\rho_0}\right)^{7/3} - \left(\frac{\rho}{\rho_0}\right)^{5/3}\right]\left\{1 + \frac{3}{4}(K_0' - 4)\left[\left(\frac{\rho}{\rho_0}\right)^{2/3} - 1\right]\right\} + p_{\mathrm{th}} \tag{33}$$

Since internal temperatures of the terrestrial planets usually exceed the Debye temperature, the quasi-harmonic approximation is valid [92And] and the thermal pressure given by

$$p_{\mathrm{th}} = \int_{T_0}^{T}\left(\frac{\partial p}{\partial T}\right)_V dT = \int_{T_0}^{T} \alpha K_T dT \approx \alpha_0 K_{0T}(T - T_0) \tag{34}$$

can be used in Eq. (33) to include the temperature effect in the equation-of-state parameters [84And]. We refer the reader to [07Soh] for a discussion of the 4th order isothermal Birch-Murnaghan equation of state applied to terrestrial planet interiors, involving second pressure and first temperature derivatives of the isothermal bulk modulus. Equation-of-state parameters of selected Mg- and Fe-rich mantle mineralogical phase assemblages and core constituents are given in Tables 2 – 5.

In summary, the simplicity of the Williamsom-Adams method is partly counterbalanced by the disadvantage that seismological data are not frequently available for terrestrial planets other than the Earth and the Moon. The linear pressure dependence of K_{T} inherent in Murnaghan's equation may represent an adequate assumption for small bodies like the Moon and the satellites of the outer giant planets [07Hus]. However, even for planets in the mass and size range of Mercury or Mars, with central pressures of the order of the isothermal bulk modulus, more reliable higher-order finite-strain equations of state corrected for thermal pressure effects such as Eq. 33 are recommended for the construction of models of the internal density distribution.

Table 2. Physical properties of mantle mineral phase assemblages.

Phase	Formula	ϱ_0 [g cm^{-3}]	$\bar{\mu}$ [g atom^{-1}]	$\theta_{\mathrm{D,Mg}}$ [K]	$\theta_{\mathrm{D,Fe}}$ [K]
Olivine (α)	$(Mg,Fe)_2SiO_4$	$3.227 + 1.175\chi_{Fe}$	$20.099 + 9.011\chi_{Fe}$	809(1)	619(2)
Wadsleyite (β)	$(Mg,Fe)_2SiO_4$	$3.472 + 1.289\chi_{Fe}$	$20.099 + 9.011\chi_{Fe}$	849(4)	656(27)
Ringwoodite (γ)	$(Mg,Fe)_2SiO_4$	$3.563 + 1.305\chi_{Fe}$	$20.099 + 9.011\chi_{Fe}$	889(6)	648(8)
Orthopyroxene	$(Mg,Fe)SiO_3$	$3.203 + 0.798\chi_{Fe}$	$20.078 + 6.308\chi_{Fe}$	809(8)	653(14)
Ca-clinopyroxene	$Ca(Mg,Fe)Si_2O_6$	$3.279 + 0.377\chi_{Fe}$	$21.655 + 3.154\chi_{Fe}$	782(5)	702(4)
Clinopyroxene HP	$(Mg,Fe)SiO_3$	$3.304 + 0.832\chi_{Fe}$	$20.078 + 6.308\chi_{Fe}$	824(7)	672(11)
Garnet	$(Mg,Fe)_3Al_2Si_3O_{12}$	$3.565 + 0.747\chi_{Fe}$	$20.156 + 4.731\chi_{Fe}$	823(4)	742(5)
Ca-perovskite	$CaSiO_3$	4.232	23.232	725(28)	–
Perovskite	$(Mg,Fe)SiO_3$	$4.106 + 1.088\chi_{Fe}$	$20.078 + 6.308\chi_{Fe}$	888(4)	700(10)
Magnesiowüstite	$(Mg,Fe)O$	$3.586 + 2.274\chi_{Fe}$	$20.152 + 15.770\chi_{Fe}$	773(9)	417(7)
Post-perovskite	$MgSiO_3$	4.071^a	20.078	1100^a	–

Note: ϱ_0, specific mass, $\bar{\mu}$, mean atomic mass, θ_{D}, Debye temperature, χ_{Fe}, mole fraction of iron. All values are taken at standard ambient pressure and room-temperature conditions (300 K). Indices Mg and Fe refer to the specific mineral endmember species. Error estimates in the last digits are given in parentheses.

References: Unless stated otherwise, the data to calculate specific mass and Debye temperature were taken from [05Sti, 07Sti]. A compilation of atomic weights by [06Wie2] was used to calculate mean atomic mass. $^{a)}$ [04Tsu].

Table 3. Elastic properties of mantle mineral phase assemblages.

Phase	$V_{0,\mathrm{Mg}}$ [cm^3 mol^{-1}]	$V_{0,\mathrm{Fe}}$ [cm^3 mol^{-1}]	$K_{0\mathrm{T,Mg}}$ [GPa]	$K_{0\mathrm{T,Fe}}$ [GPa]	$K'_{0,\mathrm{Mg}}$	$K'_{0,\mathrm{Fe}}$	$\gamma_{0,\mathrm{Mg}}$	$\gamma_{0,\mathrm{Fe}}$	l_{Mg}	l_{Fe}
Olivine (α-phase)	43.60	46.29	128(2)	135(2)	4.2(2)	4.2(10)	0.9(3)	1.06(7)	2.1(2)	3.6(10)
Wadsleyite (β-phase)	40.52	42.80	169(3)	169(13)	4.3(2)	4.3(10)	1.2(9)	1.21(30)	2.0(10)	2.0(10)
Ringwoodite (γ-phase)	39.49	41.86	183(2)	199(7)	4.1(2)	4.1(10)	1.0(10)	1.17(22)	2.8(4)	2.8(10)
Orthopyroxene	125.35	131.88	107(2)	100(4)	7.0(4)	7.0(5)	0.8(4)	0.75(8)	0.3(11)	0.3(10)
Ca-clinopyroxene	132.08	135.73	112(5)	119(4)	5.2(18)	5.2(10)	0.9(5)	0.93(6)	1.5(20)	1.5(10)
Clinopyroxene HP	121.52	127.59	121(1)	121(10)	5.5(3)	5.5(10)	1.1(5)	1.16(30)	0.8(5)	0.8(10)
Garnet	113.08	115.43	170(2)	177(3)	4.1(3)	4.1(3)	1.0(6)	1.10(6)	1.4(5)	1.4(10)
Ca-perovskite	27.45	–	236(4)	–	3.9(2)	–	1.4(7)	–	1.7(16)	–
Perovskite	24.45	25.40	251(3)	281(40)	4.1(1)	4.1(10)	1.4(5)	1.44(30)	1.4(5)	1.4(10)
Magnesiowüstite	11.24	12.26	161(3)	179(1)	3.9(2)	4.9(2)	1.5(2)	1.53(13)	1.5(2)	1.5(10)
Post-perovskite	24.66	–	222(1)	–	4.2(1)	–	1.64	–	1.9	–

Note: V_0, molar volume, $K_{0\mathrm{T}}$, isothermal bulk modulus, K'_0, first pressure derivative of isothermal bulk modulus, γ_0, thermodynamic Grüneisen parameter, $l = -\frac{\partial \ln \gamma}{\partial \ln \rho}$ logarithmic volume derivative of thermodynamic Grüneisen parameter. All values are taken at standard ambient pressure and room-temperature conditions (300 K). Indices Mg and Fe refer to the specific mineral endmember species. Error estimates in the last digits are given in parentheses. Isothermal bulk moduli K_T can be related to adiabatic bulk moduli K_S by using the isothermal-adiabatic transformation Eq. (29).

References: Data are taken from the compilation of [05Sti, 07Sti] except those for post-perovskite that are taken from [04Tsu].

Table 4. Thermal expansivity of mantle mineral phase assemblages.

Phase	α_0 [10^{-5} K^{-1}]	α_1 [10^{-8} K^{-2}]	α_2 [K]
Olivine (α-phase)[a]	2.832	0.758	0.
Wadsleyite (β-phase)[b]	2.711	0.6885	0.5767
Ringwoodite (γ-phase)[c]	1.872	0.421	0.6537
Orthopyroxene[d]	2.86	0.72	0.
Ca-clinopyroxene[e]	2.32	1.88	0.
Clinopyroxene HP[d]	2.86	0.72	0.
Garnet[f]	2.81	0.316	0.4587
Ca-perovskite[g]	3.133	0.388	0.
Perovskite[h]	2.461	0.165	0.
Magnesiowüstite[i]	3.000	1.200	0.
Post-perovskite[j]	2.57	0.	0.

Note: α_0, α_1, α_2, are coefficients to calculate thermal expansivity according to $\alpha(T) = \alpha_0 + \alpha_1 T - \alpha_2 T^{-2}$, where T is expressed in units of K. Coefficients are compiled for Mg-type mineral endmember species.

References: [a)] [96Bou]; [b)] [92Fei]; [c)] [94Men]; [d)] [95Zha]; [e)] [98Zha]; [f)] [91And]; [g)] [96Wan, 05Mat]; [h)] [98Fiq, 00Fiq, 05Mat]; [i)] [00Dew]; [j)] [05Tsu].

Table 5. Thermodynamic properties of core constituents.

Phase	ϱ_0 [kg m^{-3}]	$K_{0\mathrm{T}}$ [GPa]	K'_0	γ_0	α [10^{-5} K^{-1}]	θ_D [K]
α-Fe[a]	7874	167	4.01	1.78[h]	6.93[h]	430[h]
γ-Fe[b]	8200	155	5.5	1.78[h]	6.93[h]	430[h]
ϵ-Fe[c]	8298	165	5.33	1.78[i]	6.93[m]	430[i]
Fe (liq.)[d]	8000	180	4.66	1.74[j]	9.2[j]	-[o]
FeS IV[e]	5100	60	4.0	1.54[k]	6.85[n]	674[q]
FeS VI[f]	5869	156	4.0	1.54[l]	-[o]	674[l]
Fe + 10 wt.% S (liq.)[g]	6400	118	4.7	1.40[k]	9.2[p]	-[o]

Note: ϱ_0, specific mass, K_{0T}, isothermal bulk modulus, K'_0, first pressure derivative of isothermal bulk modulus, γ_0, thermodynamic Grüneisen parameter, α, thermal expansivity, θ_D Debye temperature. All values are taken at standard ambient pressure and room-temperature conditions (300 K). Isothermal bulk moduli K_T can be related to adiabatic bulk moduli K_S by using the isothermal-adiabatic transformation Eq. (29).

References: [a)] [00Dub1]; [b)] [02And, 89Boe, 90Boe]; [c)] [90Mao]; [d)] [94And, 08Rin]; [e)] [95Fei, 08Rin]; [f)] [06Ono]; [g)] [03Bal, 00San, 08Rin]; [h)] same as for ϵ-Fe assumed; [i)] [00Dub2]; [j)] [94And], α is valid in the high-temperature regime (1800 K); [k)] [84Bro]; [l)] same as for FeS IV assumed; [m)] [00Dub3], α restricted to pressure- and temperature-range of 0-300 GPa and 300-1300 K, respectively; [n)] [95Fei], α valid in the low-temperature regime (800 K); [o)] not available; [p)] same as for Fe (liquid) assumed; [q)] [89Sve]

4.2.3.2.3 Typical structure models

4.2.3.2.3.1 Overview

In the following sections, fundamental traits of the interiors of the terrestrial planets and a number of satellites of the giant planets are compared. Physical parameters used for the construction of two- and three-layer structural models of the terrestrial planets are collected in Table 6.

Table 6. Physical parameters of interior structure models of the terrestrial planets.

Object	Moon	Mercury	Mars	Venus
Mass [10^{23} kg]	0.73477	3.3019	6.4185	48.685
Radius [km]	1737.4	2439.7	3389.5	6051.8
Mean density [kg m^{-3}]	3344.7	5427.0	3934.9	5243.9
Mean moment-of-inertia factor	0.3931[a]	0.336[b]	0.3654[c]	0.33[d]
Average crust thickness [km]	45	none	45	35
Average crust density [kg m^{-3}]	2800	none	2900	2900
Average core temperature [K]	1700[d]	1700[d]	1700[d]	2800[d]
Core constituents	α-Fe/FeS IV	γ-Fe/FeS IV	γ-Fe/FeS IV	ϵ-Fe/FeS VI

Note: Whereas two-layer structural models lacking a crust are calculated for Mercury, the construction of three-layer structural models of the Moon, Mars and Venus involves average crustal thickness and density values taken from [07Wie].

References: [a] [01Kon]; [b] [01Spo]; [c] [06Kon]; [d] value assumed.

4.2.3.2.3.2 Mercury

Mercury is the smallest among the terrestrial planets with a mean radius of 2439.7 ± 1 km [07Sei]. Its mean density of 5430 ± 10 kg m^{-3} [87And], much larger than that of the Moon and Mars, is close to that of the Earth and Venus, indicating that iron is more abundant than in any other terrestrial planet. If Mercury were fully differentiated with a core mainly composed of iron, the core radius would occupy up to about 3/4 of the radius of the planet [03Sol]. The possible existence of a self-sustained magnetic field can be attributed to precipitation of a solid inner core, provided volatile constituents such as sulfur were available to keep the outer core liquid, prerequisite to core dynamo action [88Sch, 02Bal].

Doppler radio tracking data of the Mariner 10 and MESSENGER spacecraft provided information on the planet's mass ($GM = 22,032.09 \pm 0.91$ km^3 s^{-2}, G is the gravitational constant). The mean radius of Mercury has been derived from Earth-based radar ranging and Mariner 10 range observations [87And]. Mercury's large 88-day forced libration amplitude in longitude of $35.8'' \pm 2''$ and Cassini-state-type obliquity of $2.11' \pm 0.2'$ as obtained from ground-based radar interferometry observations [07Mar] together with the second-degree-and-order gravitational coefficient $C_{2,2} = (1 \pm 0.5) \times 10^{-5}$ obtained from Mariner 10 observations [87And] support the view that the silicate mantle of the planet is mechanically decoupled from a partially molten core.

Based on gravity and shape data acquired during consecutive Mariner 10 flybys, [96And1] estimated crustal thicknesses in the range from 100 to 300 km which were difficult to reconcile with the magmatic history of the planet. [02Nim] consider the stability of Mercury's long-wavelength topography with respect to lower crustal flow, the latter of which limiting crust thickness. It has

Table 7. Two-layer structural model determinations of Mercury (mean MoI factor: 0.336).

χ_S	χ_{FeS}	ρ_c [kg m^{-3}]	ρ_m [kg m^{-3}]	R_c/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]
Thermal pressure correction included.							
0.365	1.0	6271	–	–	–	–	–
0.227	0.621	7000	–	–	–	–	–
0.186	0.508	7250	1700.7	0.8758	0.8972	35.592	2.0564
0.147	0.403	7500	2591.6	0.8330	0.7985	36.627	4.1630
0.111	0.305	7750	3047.0	0.7970	0.7229	37.629	5.8903
0.078	0.213	8000	3324.1	0.7663	0.6632	38.605	7.3422
0.046	0.126	8250	3510.9	0.7396	0.6149	39.558	8.5870
0.016	0.045	8500	3645.5	0.7161	0.5751	40.492	9.6715
0	0	8645	3707.7	0.7037	0.5550	41.027	10.241
Thermal pressure not accounted for.							
0.365	1.0	6502	–	–	–	–	–
0.236	0.647	7250	–	–	–	–	–
0.164	0.489	7750	2451.7	0.8251	0.8021	38.185	4.1684
0.100	0.275	8250	3134.7	0.7654	0.6815	40.218	7.0518
0.044	0.121	8750	3460.4	0.7192	0.5997	42.174	9.2310
0	0	9186	3631.9	0.6864	0.5473	43.834	10.761

Note: The following assumptions are made: (a) The core is composed of γ-Fe and FeS IV whose elastic parameters are taken from [89Boe, 90Boe], [02And] and [95Fei], respectively. (b) The average core density is obtained by extrapolating the third-order Birch-Murnaghan equation Eq. (33) neglecting the thermal pressure contribution to central pressure conditions of a homogeneous, equivalent-mass sphere of Mercury. (c) The thermal pressure correction at a core temperature of 1700 K may vary between 5.8 GPa (pure FeS IV core) and 15 GPa (pure γ-Fe core) according to Eq. (34). For core sulfur contents exceeding about 20 wt.%, the resultant size of the core would be as large as that of the planet.

been suggested that the maximum possible crustal thickness may range between 100 and 200 km, depending on the assumed rheology and spatial distribution of internal heat sources.

See [07vHo] and [07Soh] for more complete accounts on the Mercurian interior and Table 7 for a comparison of two-layer structural models with variable core sulfur contents χ_S.

4.2.3.2.3.3 Venus

The small size difference between Venus (mean radius 6051.8 ± 1 km) and the Earth (mean radius 6371.00 ± 0.01 km) [07Sei] may have caused completely different traits of evolution of both planets that manifest themselves in the lack of plate tectonics on Venus in contrast to the Earth. Since the moment of inertia of Venus is not known, there is no firm constraint on the planet's internal mass distribution. It is less likely that the axial moment of inertia of Venus could be derived from observations of its rotational state alone since the planet's retrograde rotation is extremely slow (rotation period of 243.0185 ± 0.001 d) and its rotation axis is almost perpendicular aligned to the plane of its nearly circular orbit [97Yod]. Nevertheless, Venus is sufficiently large that differentiation into an Earth-like structure consisting of a metallic core, silicate mantle, and basaltic crust likely accompanied hot accretion. As a consequence of the slightly smaller size and mass of Venus compared to the Earth, pressure-induced mineral phase transitions may occur at larger depths in the mantle. If the core were to contain less light elements than the Earth's core, however, the entire thickness of the Venusian mantle could even supersede that of the Earth.

Doppler radio tracking data of the Magellan and Pioneer Venus spacecraft provided important

Table 8. Three-layer structural model determinations of Venus (mean MoI factor: 0.33).

χ_S	χ_{FeS}	ρ_c [kg m^{-3}]	ρ_m [kg m^{-3}]	R_c/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]
Thermal pressure correction included.							
0.365	1.0	8388	3353.0	0.7225	0.6032	240.37	52.399
0.305	0.837	8750	3529.0	0.6914	0.5515	248.23	60.896
0.231	0.632	9250	3687.3	0.6560	0.4979	258.92	70.481
0.164	0.448	9750	3793.1	0.6267	0.4576	269.44	78.349
0.103	0.283	10250	3869.1	0.6019	0.4263	279.83	85.006
0.048	0.132	10750	3926.5	0.5805	0.4011	290.11	90.767
0	0	11232	3970.1	0.5625	0.3811	299.95	95.668
Thermal pressure not accounted for.							
0.365	1.0	8775	3538.8	0.6895	0.5484	248.77	61.427
0.290	0.794	9250	3687.3	0.6560	0.4979	258.92	70.481
0.218	0.598	9750	3793.1	0.6267	0.4576	269.44	78.349
0.154	0.422	10250	3869.1	0.6019	0.4263	279.83	85.006
0.096	0.262	10750	3926.5	0.5805	0.4011	290.11	90.767
0.043	0.117	11250	3971.6	0.5618	0.3805	300.31	95.841
0	0	11685	4003.6	0.5473	0.3653	309.14	99.809

Note: The following assumptions are made: (a) The core is composed of ϵ-Fe and FeS VI whose elastic parameters are taken from [90Mao] and [06Ono, 08Ono], respectively. (b) The average core density is obtained by extrapolating the third-order Birch-Murnaghan equation Eq. (33) neglecting the thermal pressure contribution to central pressure conditions of a homogeneous, equivalent-mass sphere of Venus. (c) The thermal pressure correction at a core temperature of 2800 K may vary between 27 GPa (pure FeS VI core) and 29 GPa (pure ϵ-Fe core) according to Eq. (34). (d) The average crust thickness and density is 35 km and 2900 kg m^{-3}, respectively, resulting in a calculated crust-mantle transition pressure of around 900 MPa.

information on the planet's mass ($GM = 324,858.6 \pm 0.014$ km^3 s^{-2}, G is the gravitational constant), its retrograde rotation state, gravitational field, and tidal Love number k_2 [97Sjo]. Crust thickness estimates range from 20 to 50 km [97Gri, 98Nim]. The crust is thicker (45-85 km) beneath the plateau highlands (Alpha, Ovda, Thetis and Tellus Regiones).

See [07Soh] for a more complete account on the Venusian interior and Table 8 for a comparison of three-layer structural models with variable core sulfur contents χ_S.

4.2.3.2.3.4 Mars

Its mean radius and density of of 3389.5 ± 0.2 km [07Sei] and 3935 kg m^{-3}, respectively, places Mars well in the midst of all other terrestrial bodies. The polar moment of inertia as derived from a combined analysis of Mars Global Surveyor tracking and Mars Pathfinder and Viking Lander ranging data indicates that the Martian interior is differentiated into an Earth-like structure consisting of a sulfuric metallic core, a silicate mantle enriched in iron oxide and overlain by a basaltic crust. Most outstanding geologic features are the approximately hemispheric dichotomy between the southern highlands and the northern plains and the near-equatorial volcanic Tharsis province that contains the largest volcanoes on Mars [92Tan].

Doppler radio tracking data of a number of orbiting and landed spacecraft have provided basic information about the planet's mass ($GM = 42,828.371901 \pm 0.000074$ km^3 s^{-2}, G is the gravitational constant [01Lem]), spin-axis precession [97Fol, 03Yod, 06Kon], tidal Love number k_2 [03Yod, 06Kon], and gravitational field [06Kon]. The global topography and gravitational field of Mars have been determined with high accuracy using laser altimetry and two-way Doppler tracking

Table 9. Three-layer structural model determinations of Mars (mean MoI factor: 0.3654).

χ_S	χ_{FeS}	ρ_c [kg m^{-3}]	ρ_m [kg m^{-3}]	R_c/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]
Thermal pressure correction included.							
0.365	1.0	6288	3553.9	0.5299	0.2377	36.416	18.592
0.274	0.750	6750	3576.6	0.4950	0.2081	37.826	19.900
0.188	0.516	7250	3592.9	0.4657	0.1860	39.302	21.003
0.114	0.311	7750	3604.4	0.4421	0.1702	40.742	21.894
0.048	0.132	8250	3613.0	0.4226	0.1582	42.157	22.642
0	0	8660	3618.6	0.4089	0.1504	43.304	23.176
Thermal pressure not accounted for.							
0.365	1.0	6518	3566.5	0.5114	0.2215	37.124	19.287
0.279	0.764	7000	3585.5	0.4795	0.1961	38.569	20.484
0.201	0.551	7500	3599.1	0.4533	0.1775	40.025	21.470
0.133	0.364	8000	3609.0	0.4319	0.1638	41.452	22.283
0.073	0.200	8500	3616.6	0.4140	0.1533	42.858	22.974
0	0	9197	3624.6	0.3932	0.1421	44.793	23.793

Note: The following assumptions are made: (a) The core is composed of γ-Fe and FeS IV whose thermodynamic parameters are taken from [89Boe, 90Boe], [02And], and [95Fei], respectively. (b) The average core density is obtained by extrapolating the third-order Birch-Murnaghan equation Eq. (33) neglecting the thermal pressure contribution to central pressure conditions of a homogeneous, equivalent-mass sphere of Mars. (c) The thermal pressure correction at a core temperature of 1700 K may vary between 5.8 GPa (pure FeS IV core) and 15 GPa (pure γ-Fe core) according to Eq. (34). (d) The average crust thickness and density is 45 km and 2900 kg m^{-3}, respectively, resulting in a calculated crust-mantle transition pressure of around 485 MPa.

of the Mars Global Surveyor spacecraft [99Smi1, 99Smi2, 00Zub, 01Lem]. The re-analysis of the entire data set resulted in an improved value of the polar moment-of-inertia factor of $C/M/a^2 = 0.3654 \pm 0.0008$ with the *equatorial* radius a fixed at 3396 km [06Kon], being consistent with the model of a mostly hydrostatic planet and non-hydrostatic contributions to the MoI factor entirely related to the axisymmetric distribution of topographic loads about Tharsis [79Kau]. Taking into account the planet's gravitational oblateness and minor contributions to oblateness due to the Tharsis rise would result in a mean moment-of-interia $I/M/R^2 \approx 0.3654 \pm 0.0008$, suggesting a slightly stronger concentration of mass toward the center than previously thought, with consequences for the planet's bulk chemistry and interior structure.

Present estimates of the mean crustal thickness of Mars are entirely based on indirect geophysical evidence, geochemical arguments, and arbitrarily assumed crust and mantle densities and, therefore, may range between 30 and 80 km [04Neu, 04Wie, 05Sol] although crust thicknesses of up to about 100 km were also consistent with global geophysical constraints [97Soh, 01Kav, 04Gud, 05Soh].

See [07Soh] for a more complete account on the Martian interior and Table 9 for a comparison of three-layer structural models with variable core sulfur contents χ_S.

4.2.3.2.3.5 The Moon

The mean radius (1737.4 ± 1 km [07Sei]) of the Moon is about one quarter of that of the Earth. Its mean density of 3344 kg m^{-3}, however, is comparatively low, indicating that iron is much less abundant than in the Earth or any other terrestrial planetary body. If the Moon were fully differentiated with a pure iron core, the core radius would occupy less than 1/4 of the lunar radius. Lunar laser ranging data show that the true spin axis of the Moon is displaced from the

Table 10. Two-layer structural model determinations of the Moon (mean MoI factor: 0.3931).

χ_S	χ_{FeS}	ρ_c [kg m^{-3}]	ρ_m [kg m^{-3}]	R_c/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]
Thermal pressure correction included.							
0.365	1.0	5010	3279.8	0.3349	0.05626	5.5735	4.3859
0.260	0.713	5500	3281.1	0.3061	0.04718	5.7112	4.5152
0.171	0.468	6000	3282.0	0.2848	0.04144	5.8436	4.6119
0.095	0.261	6500	3282.6	0.2684	0.03756	5.9712	4.6876
0.030	0.083	7000	3283.0	0.2551	0.03475	6.0953	4.7500
0	0	7262	3283.2	0.2491	0.03357	6.1594	4.7787
Thermal pressure not accounted for.							
0.365	1.0	5451	3281.0	0.3086	0.04789	5.6978	4.5042
0.307	0.841	5750	3281.6	0.2947	0.04401	5.7782	4.5668
0.222	0.608	6250	3282.3	0.2761	0.03933	5.9079	4.6518
0.150	0.410	6750	3282.8	0.2614	0.03605	6.0335	4.7202
0.087	0.239	7250	3283.2	0.2494	0.03362	6.1565	4.7775
0.033	0.090	7750	3283.6	0.2393	0.03175	6.2775	4.8267
0	0	8088	3283.7	0.2333	0.03071	6.3588	4.8566

Note: The following assumptions are made: (a) The core is composed of α-Fe and FeS IV whose thermodynamic parameters are taken from [00Dub1] and [95Fei], respectively. (b) The average core density is obtained by extrapolating the third-order Birch-Murnaghan equation Eq. (33) neglecting the thermal pressure contribution to central pressure conditions of a homogeneous, equivalent-mass sphere of the Moon. (c) The thermal pressure correction at a core temperature of 1700 K may vary between 5.8 GPa (pure FeS IV core) and 16 GPa (pure α-Fe core) according to Eq. (34).

Cassini alignment (mean direction of the spin axis) by $0.26''$, indicating internal dissipation in the presence of a fluid core, not necessarily composed of iron [81Yod, 94Dic, 01Wil]. Furthermore, remnant magnetization of the lunar crust and the paleomagnetic record of some lunar samples provide circumstantial evidence for the existence of a lunar core, provided that magnetization was acquired from an intrinsic magnetic field caused by an early lunar core dynamo [87Hoo].

From Doppler tracking of the Lunar Prospector spacecraft and lunar laser ranging data on lunar libration, [01Kon] have obtained basic information on the gravitational field of the Moon and improved values of the Moon's mass ($GM = 4902.801076 \pm 0.000081$ km^3 s^{-2}, G is the gravitational constant), mean moment-of-inertia factor ($I/M/R^2 = 0.3931 \pm 0.0002$), and tidal potential Love number ($k_2 = 0.026 \pm 0.003$). The MoI factor is consistent with a core radius between 220 and 450 km. Lunar Prospector observations of the induced lunar magnetic moment when the Moon passes through the geomagnetic tail lobes of the Earth are consistent with a lunar core radius of 340 ± 90 km [99Hoo]. The existence of a small lunar core could not be confirmed or rejected on the basis of the seismic data acquired by the Apollo missions.

From a reanalysis of gravity data acquired by the Apollo and Clementine spacecraft and assuming a thickness of 55 km at the Apollo 12/14 site, the thickness of the lunar farside crust has been estimated at 67 km. A mean thickness of 61 km is obtained if a uniform crustal composition is assumed [96Neu]. The mean thickness of the lunar crust is estimated at 49 ± 15 km if an Airy-type compensation mechanism applies [06Wie1, 07Hik].

See [06Wie1] and [07Soh] for more complete accounts on the lunar interior and Table 10 for a comparison of two-layer structural models with variable core sulfur contents χ_S.

4.2.3.2.3.6 The Galilean satellites

Internal structure models of the Galilean Satellites, Io, Europa, Ganymede, and Callisto, have been constructed from data on the satellites' gravity fields and shapes acquired by the *Galileo* spacecraft in orbit around Jupiter from 1995–2003. The mission included several close flybys of the moons and the analysis of gravitational perturbations of the spacecraft allowed the determination of the principal moment of inertia of the satellites, a quantity which is indicative of a satellite's concentration of mass towards the center. However, the moment-of-inertia (MoI) values were derived assuming that the satellites are in hydrostatic equilibrium, an assumption that could not be verified from the data. An exception is Io, for which hydrostatic equilibrium was confirmed from different flyby geometries (near-polar and near-equatorial). The models described here do assume hydrostatic equilibrium and will neglect self-compression of chemically uniform layers because of sufficiently low internal pressures [02Soh]. The main data which have been used to construct these models are collected in Tables 11 and 12.

Table 11. Physical parameters of the Galilean Satellites.

satellite	R [km]	GM [km^{3}s^{-2}]	$\bar{\rho}$ [kg m^{-3}]	a [km]	b [km]	c [km]
Io	1821.46	5959.91 ± 0.02	3529	1829.4	1819.3	1815.7
Europa	1562.09	3202.72 ± 0.02	3006	1564.13	1561.23	1560.93
Ganymede	2632.345	9887.83 ± 0.03	1940	2632.4	2632.29	2632.35
Callisto	2409.3	7179.29 ± 0.01	1837	2409.4	2409.2	2409.3

Note: The mean densities are calculated from $\bar{\rho} = 3M/(4\pi R^3)$. Values of mean radius R and mass GM (G is the gravitational constant) were taken from [07Sei] and [04Sch]. a, b, and c denote the radius along the Jupiter-facing, orbit-facing, and polar axis, respectively.

Table 12. Gravity field parameters of the Galilean satellites. $R_{\rm ref}$ is the reference radius associated with J_2 and $C_{2,2}$; the corresponding values of GM are given in Table 11. J_2 ($= -C_{2,0}$) and $C_{2,2}$ are low-order gravity coefficients. $k_{\rm f}$ is the fluid potential Love number, and $C/M/R^2$ is the dimensionless MoI factor with respect to the rotational axis.

satellite	$R_{\rm ref}$ [km]	$J_2 \times 10^{-6}$	$C_{2,2} \times 10^{-6}$	$k_{\rm f}$	$C/(MR^2)$
Io[a]	1821.6 ± 0.5	1859.5 ± 2.7	558.8 ± 0.8	1.3043 ± 0.0019	0.37824 ± 0.00022
Europa[a]	1565.0 ± 8.0	435.5 ± 8.2	131.5 ± 2.5	1.048 ± 0.020	0.346 ± 0.005
Ganymede[a]	2631.2 ± 1.7	127.53 ± 2.9	38.26 ± 0.87	0.804 ± 0.018	0.3115 ± 0.0028
Callisto[a]	2410.3 ± 1.5	32.7 ± 0.8	10.2 ± 0.3	1.103 ± 0.035	0.3549 ± 0.0042

Note: The static fluid potential Love numbers are calculated from $k_{\rm f} = 4C_{2,2}/q_r$, where $q_r = \omega^2 R^3_{\rm ref}/(GM)$ is, to first order, the smallness parameter for the equilibrium figure of a synchronously rotating satellite with ω the mean angular frequency of the rotational and orbital period.
Reference: [a)] [04Sch]

Io's mean density of 3529 kg m^{-3} suggests that this satellite is composed of silicates and iron. Galileo Doppler radio tracking data have been used to calculate the satellite's MoI-factor from a determination of the tidally and rotationally induced oblateness (J_2) and the equatorial ellipticity ($C_{2,2}$) of its gravitational field. Interior structure models matching the best estimate of Io's MoI factor of 0.37824 ± 0.00022 give iron and eutectic Fe-FeS core mass fractions of 10.5% and 20.2%, respectively [96And2]. The independent determination of the quadrupole coefficients J_2 and $C_{2,2}$

Table 13. Two-layer structural model determinations of Io (mean MoI factor: 0.3782).

ρ_c [kg m^{-3}]	ρ_m [kg m^{-3}]	R_c/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]
5000	3258.0	0.5375	0.2201	7.5518	4.2028
5400	3274.5	0.4926	0.1829	7.8406	4.5596
5800	3284.6	0.4595	0.1594	8.1131	4.8204
6200	3291.5	0.4336	0.1432	8.3748	5.0241
6600	3296.5	0.4126	0.1314	8.6289	5.1906
7000	3300.4	0.3951	0.1224	8.8777	5.3311
7400	3303.4	0.3802	0.1152	9.1225	5.4526
7800	3305.9	0.3673	0.1095	9.3645	5.5597

Table 14. Three-layer structural model determinations of Europa (mean MoI factor: 0.346).

ρ_c [kg m^{-3}]	ρ_m [kg m^{-3}]	ρ_s [kg m^{-3}]	R_c/R	R_m/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]	p_{base} [MPa]
5150	2800	1000	0.5592	61.828	0.2996	5.1270	2.2992	82.827
5150	2900	1000	0.5437	73.644	0.2755	5.0939	2.4199	99.070
5150	3000	1000	0.5268	84.968	0.2506	5.0582	2.5479	114.77
5150	3100	1000	0.5081	95.872	0.2247	5.0191	2.6845	130.01
5150	3200	1000	0.4869	106.42	0.1978	4.9755	2.8311	144.87
5150	3300	1000	0.4627	116.67	0.1697	4.9259	2.9897	159.44
5150	3400	1000	0.4342	126.68	0.1403	4.8679	3.1627	173.78
5150	3500	1000	0.3995	136.51	0.1093	4.7975	3.3538	187.96
5150	3600	1000	0.3546	146.16	0.0764	4.7059	3.5688	202.01
5150	*3700*	*1000*	*0.2906*	*155.84*	*0.0420*	*4.5784*	*3.8147*	*216.22*
5150	*3800*	*1000*	*0.1497*	*165.52*	*0.0057*	*4.3240*	*4.1214*	*230.54*

Note: The following assumptions are made: (a) The density of the Fe-FeS core ρ_c is fixed at a value corresponding to the eutectic sulfur concentration at core pressures. (b) The density of the water-ice/liquid outer shell ρ_s is taken constant, i.e. a small density contrast between the outer ice shell and the underlain water ocean is neglected. (c) The possible range of meaningful models is limited by the pressure-induced transition from ice I to ice II at around 209 MPa. Emphasized in italic are model determinations superseding that critical pressure at the base of the outer ice shell.

using three Galileo flybys is consistent with Io being in hydrostatic equilibrium [01And]. However, deviations from hydrostatic equilibrium due to, e.g., a vigorously convecting interior, may exist at an undetectable level.

See Table 13 for a comparison of two-layer structural models of Io with variable core composition and density.

The mean density of **Europa** of 3006 kg m^{-3} is intermediate between those of icy satellites such as Ganymede and Callisto and rocky satellites such as Io and the Moon. Voyager and Galileo imaging and infrared spectroscopy showed that Europa's surface is covered with water ice which together with the density calls for models that range from hydrated rock-metal interiors beneath relatively thin ice shells to dehydrated interiors underneath thick ice shells. However, radiogenic and tidal heating is thought to have been sufficient for dehydration and even subsequent differentiation of the rock-metal component to form an iron-rich core overlain by a silicate mantle [98And1].

Particularly, strike-slip behavior and plate rotation near the anti-Jovian point is taken as the current best geologic evidence that the water ice shell is decoupled from Europa's deep interior due to the existence of a subsurface liquid water ocean or at least a soft ice layer [99Pap]. The most convincing argument for an ocean results from an interpretation of Galileo magnetometer data [00Kiv] that requires an electrically conducting layer at a shallow depth in which a magnetic field is induced as Europa moves through the magnetosphere of Jupiter. It is possible that the ice layer itself is tidally heated at a rate sufficient to stabilize the ice layer against freezing [02Hus], but [03Spo] have shown that even radiogenic heating in a chondritic core will suffice to keep a subsurface water ocean.

See Table 14 for a comparison of three-layer structural models of Europa with variable silicate mantle density.

Ganymede is the largest satellite in the solar system and exceeds Mercury in diameter. Gravitational and magnetic field observations by the Galileo spacecraft together with spectral data of the surface suggest that Ganymede is strongly differentiated [04Sch]. The spectral data are consistent with a surface mainly covered by water ice. The mean density of 1940 kg m^{-3} indicates that the interior is composed of water ice and rock-metal components in nearly equal amounts by mass.

Table 15. Three-layer structural model determinations of Ganymede (mean MoI factor: 0.3115).

ρ_c [kg m^{-3}]	ρ_m [kg m^{-3}]	ρ_s [kg m^{-3}]	R_c/R	R_m/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]	p_{base} [GPa]
5600	2800	1200	0.3954	0.7071	0.1786	9.8132	5.0645	1.4082
5600	2900	1200	0.3828	0.7018	0.1620	9.7152	5.2650	1.4376
5600	3000	1200	0.3732	0.6948	0.1501	9.6730	5.4434	1.4755
5600	3100	1200	0.3611	0.6889	0.1360	9.6029	5.6429	1.5081
5600	3200	1200	0.3472	0.6835	0.1209	9.5188	5.8572	1.5385
5600	3300	1200	0.3329	0.6779	0.1065	9.4412	6.0763	1.5698
5600	3400	1200	0.3157	0.6727	0.0909	9.3419	6.3158	1.5989
5600	3500	1200	0.2957	0.6677	0.0746	9.2274	6.5728	1.6274
5600	3600	1200	0.2716	0.6628	0.0578	9.0905	6.8509	1.6554
5600	3700	1200	0.2409	0.6581	0.0404	8.9183	7.1553	1.6830
5600	3800	1200	0.1976	0.6533	0.0223	8.6801	7.4942	1.7105

Note: The following assumptions are made: (a) The density of the Fe-FeS core ρ_c is fixed at a value corresponding to the eutectic sulfur concentration at core pressures. (b) The density of the water-ice/liquid outer shell ρ_s is taken constant at an average value taking into account the possible presence of high-pressure phases such as ice V and ice VI. (c) The possible range of meaningful models is limited by the most likely range of silicate mantle densities.

The MoI factor of 0.3115 ± 0.0028 is the smallest measured value for any solid body in the solar system and indicates a strong concentration of mass towards the center of Ganymede [96And3]. Magnetometer measurements of the Galileo spacecraft have shown that Ganymede possesses an intrinsic magnetic field with equatorial and polar field strengths at the surface of 750 and 1200 nT, respectively. Since the most likely source is dynamo action in a liquid Fe-FeS core [96Schu], Ganymede's interior is believed to consist of an iron-rich core surrounded by a silicate rock mantle overlain by an ice shell that may contain a subsurface water ocean sandwiched between a high-pressure water ice layer and an outermost ice I layer [96And3, 02Soh].

See Table 15 for a comparison of three-layer structural models of Ganymede with variable mantle density.

Callisto's radius is about 200 km smaller than that of Ganymede and its mass is 70% that of Ganymede, resulting in a mean density of 1837 kg m^{-3} and implying that the interior is likewise composed of water ice and rock-metal components in nearly equal amounts by mass. The satellite's

Table 16. Two-layer structural model determinations of Callisto (mean MoI factor: 0.3549).

ρ_m [kg m^{-3}]	ρ_c [kg m^{-3}]	R_c/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]
900	2262.9	0.8826	0.8469	3.5640	0.3274
1000	2282.1	0.8675	0.8110	3.5885	0.4083
1100	2308.9	0.8479	0.7662	3.6220	0.5119
1200	2349.0	0.8215	0.7089	3.6710	0.6495
1300	2415.3	0.7838	0.6330	3.7494	0.8415
1400	2545.4	0.7253	0.5287	3.8953	1.1298
1500	2908.5	0.6208	0.3788	4.2686	1.6232

old, heavily cratered, and dirt-rich surface suggests that endogenic resurfacing has never happened since it completed its accretion. Provided hydrostatic equilibrium is attained, the Galileo gravity data suggest that the satellite's MoI axial factor is equal to 0.3549 ± 0.0042. However, this value is not compatible with a fully differentiated interior of Callisto and suggests partial internal differentiation [98And2, 01And2], augmented by a density increase with depth due to pressure-induced water ice phase transitions [97Kin].

Possible two-layer model determinations are summarized in Table 16 to illustrate the effect of variable degree of partial differentiation on Callisto's internal density distribution. The magnetic data suggest that an ocean is present at depth around 150 km [98Khu, 00Zim]. These two interpretations of geophysical data seem contradictory since the presence of an ocean would lead to differentiation. In order to reconcile these two observations, [04Nag] propose that Callisto's core is a mixture of ice and rock with the rock concentration increasing with depth and near the close-packing limit.

See Table 17 for a comparison of three-layer structural models of Callisto with variable mantle density.

Table 17. Three-layer structural model determinations of Callisto (mean MoI factor: 0.3549).

ρ_c [kg m^{-3}]	ρ_m [kg m^{-3}]	ρ_s [kg m^{-3}]	R_c/R	R_m/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]	p_{base} [MPa]
3500	1550	1000	0.5280	0.9994	0.2806	4.8207	2.0499	1.6666
3500	1600	1000	0.5161	0.9859	0.2620	4.7801	2.1337	41.973
3500	1650	1000	0.5036	0.9736	0.2435	4.7390	2.2188	79.004
3500	1700	1000	0.4906	0.9622	0.2250	4.6970	2.3055	113.32
3500	1750	1000	0.4768	0.9517	0.2066	4.6537	2.3943	145.36
3500	1800	1000	0.4622	0.9418	0.1882	4.6084	2.4854	175.47
3500	1850	1000	0.4465	0.9325	0.1697	4.5607	2.5794	203.95
3500	*1900*	*1000*	*0.4295*	*0.9237*	*0.1510*	*4.5097*	*2.6764*	*231.01*
3500	*1950*	*1000*	*0.4109*	*0.9154*	*0.1322*	*4.4546*	*2.7771*	*256.87*

Note: The following assumptions are made: (a) The density of the satellite's central portion ρ_c is taken constant at a value corresponding to the mean density of Io to allow for the presence of a rock-metal component. (b) The density of the water-ice/liquid outer shell ρ_s is representative for that of clean ice I separated from a mixture of high-pressure water ice and rock-metal component underneath. (c) The possible range of meaningful models is limited by the pressure-induced transition from ice I to ice II at around 209 MPa. Emphasized in italic are model determinations that supersede that critical pressure at the base of the outer ice shell.

We refer the reader to [04Sch], [07Hus] and [09Sch] for more complete accounts on the interiors of the Galilean satellites.

4.2.3.2.3.7 The Saturnian satellites

The Saturnian satellites have been investigated in detail by the *Cassini* spacecraft in orbit around Saturn since 2004. Additionally, there was the landing of the *Huygens* probe on Titan's surface in 2005 – the first touch-down on a surface of an outer solar system satellite – as part of the *Cassini/Huygens* mission.

Table 18. Physical parameters of the largest Saturnian satellites.

satellite	R [km]	GM [km^{3}s^{-2}]	$\bar{\rho}$ [kg m^{-3}]	a [km]	b [km]	c [km]
Mimas	198.2 ± 0.5	2.5023 ± 0.0020	1150 ± 9	207.4 ± 0.7	196.8 ± 0.6	190.6 ± 0.3
Enceladus	252.1 ± 0.2	7.2096 ± 0.0067	1608 ± 5	256.6 ± 0.6	251.4 ± 0.2	248.3 ± 0.2
Tethys	533.0 ± 1.4	41.2097 ± 0.0063	973 ± 8	540.4 ± 0.8	531.1 ± 2.6	527.5 ± 2.0
Dione	561.7 ± 0.9	73.1127 ± 0.0025	1476 ± 7	563.8 ± 0.9	561.0 ± 1.3	561.7 ± 0.9
Rhea	764.3 ± 2.2	153.9416 ± 0.0049	1233 ± 11	767.2 ± 2.2	762.5 ± 0.8	763.1 ± 1.1
Titan	2575.5 ± 2	8978.1356 ± 0.0039	1880 ± 4			
Iapetus	735.6 ± 3.0	120.5117 ± 0.0173	1083 ± 13	747.4 ± 3.1		712.4 ± 2.0

Note: The mean densities are calculated from $\bar{\rho} = 3M/(4\pi R^3)$. Values of mean radius R and mass GM (G is the gravitational constant) were taken from [07Tho] and [06Jac]. a, b, and c denote the radius along the Saturn-facing, orbit-facing, and polar axis, respectively.

Table 19. Gravity field parameters of selected Saturnian satellites. R_{ref} is the reference radius associated with J_2 and $C_{2,2}$; the corresponding values of GM are given in Table 18. J_2 ($= -C_{2,0}$) and $C_{2,2}$ are low-order gravity coefficients. k_{f} is the fluid potential Love number, and $C/M/R^2$ is the dimensionless MoI factor with respect to the rotational axis.

satellite	R_{ref} [km]	$J_2 \times 10^{-6}$	$C_{2,2} \times 10^{-6}$	k_{f}	$C/(MR^2)$
Rhea^a	764.4 ± 1.1	794.7 ± 89.2	235.26 ± 4.76	1.2517 ± 0.0253	0.3721 ± 0.0036
Rhea^b	764.4 ± 1.1	889 ± 25	266.6 ± 7.5	1.418 ± 0.040	0.3911 ± 0.0045
Titan	2575.5 ± 2.0	-	-	-	0.34^c

Note: The static fluid potential Love numbers are calculated from $k_{\mathrm{f}} = 4C_{2,2}/q_r$, where $q_r = \omega^2 R_{\mathrm{ref}}^3/(GM)$ is, to first order, the smallness parameter for the equilibrium figure of a synchronously rotating satellite with ω the mean angular frequency of the rotational and orbital period.

References: $^{a)}$[07Ies], based on an independent determination of J_2 and $C_{2,2}$. $^{b)}$[07And], based on a-priori hydrostatic assumption $J_2/C_{2,2} = 10/3$. $^{c)}$value assumed.

Masses, radii, and shape data of the largest Saturnian satellites are given in Table 18. Most of the icy satellites of Saturn are well-described by triaxial ellipsoids. However, Iapetus is best represented by an oblate spheroid (no difference between Saturn-facing and orbit-facing axes, i.e. $a = b$ in Table 18). For Titan all the axes are equal in length within the given uncertainties which are mainly due to the dense atmosphere enshrouding Saturn's largest moon.

The mean density and axial moment of inertia of **Rhea** as inferred from Doppler data acquired during a close Cassini spacecraft encounter on November 26, 2005 suggest that the satellite's interior

Table 20. Two-layer model structural determinations of Rhea (mean MoI factor: 0.3721).

ρ_m [kg m^{-3}]	ρ_c [kg m^{-3}]	R_c/R	M_c/M	p_c [MPa]	p_{cmb} [MPa]
800	1403.6	0.8952	0.8166	145.92	17.003
900	1421.3	0.8612	0.7364	147.32	24.968
1000	1465.0	0.7943	0.5954	150.56	39.995
1100	1733.1	0.5944	0.2953	166.86	80.190
1200	–	–	–	–	–

Table 21. Two-layer structural model determinations of Titan (mean MoI factor: 0.34).

ρ_m [kg m^{-3}]	ρ_c [kg m^{-3}]	R_c/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]
900	2530.4	0.8439	0.8090	4.7470	0.5186
1000	2570.9	0.8243	0.7661	4.8096	0.6449
1100	2628.9	0.7990	0.7134	4.8973	0.8058
1200	2718.6	0.7650	0.6475	5.0295	1.0185
1300	2874.9	0.7168	0.5632	5.2518	1.3143
1400	3211.8	0.6423	0.4526	5.7073	1.7618
1500	4401.5	0.5078	0.3066	7.2053	2.5727
1600	–	–	–	–	–
1800	–	–	–	–	–

is composed of about 25% rock-metal and 75% water ice by mass and only weakly differentiated [07Ies] or even an almost homogeneous mixture of rock-metal and water ice components [07And]. Table 20 contains possible two-layer model determinations being consistent with those gravitational field observations (Table 19), illustrating the effect of variable degree of partial differentiation on Rhea's internal density distribution.

Titan is intermediate between the Jovian satellites Ganymede and Callisto with respect to its radius of 2575.5±2 km and mean density of 1880±4 kg m^{-3}. The mean densities of Ganymede, Callisto, and Titan indicate that their interiors are composed of water ice and rock-metal component in nearly equal amounts by mass. The interior of Titan is likely to be differentiated at least into a rock-metal core and an icy mantle as a consequence of substantial accretional heating accompanied by partial outgassing and atmosphere formation [00Gras, 03Soh, 06Tob]. Whether or not Titan's deep interior is further differentiated like Ganymede's into an iron core and a rock mantle above it, is more speculative. Based on gravity data collected so far during several Cassini spacecraft encounters, it cannot be safely excluded that Titan's interior is even only partly differentiated, more similar to that of Callisto. Table 21 contains possible two-layer model determinations being consistent with preliminary inferences of the axial MoI factor from Cassini gravity data (Table 19), illustrating the effect of variable degree of partial differentiation on Titan's internal density distribution.

See [07Hus] for a more complete account on the interiors of the Saturnian satellites and Table 22 for a comparison of three-layer structural models of Titan with variable mantle density.

4.2.3.2.3.8 Icy satellites of Uranus and Neptune

The data for the Uranian satellites mainly come from ground-based observations and from the *Voyager 2* flyby in 1986. In 1989 Voyager 2 reached the Neptune system and obtained data for

Table 22. Three-layer structural model determinations of Titan (mean MoI factor: 0.34).

ρ_c [kg m^{-3}]	ρ_m [kg m^{-3}]	ρ_s [kg m^{-3}]	R_c/R	R_m/R	M_c/M	p_c [GPa]	p_{cmb} [GPa]	p_{base} [MPa]
3500	1450	1000	0.5958	0.9972	0.3938	6.0744	2.0435	9.7385
3500	1500	1000	0.5876	0.9824	0.3777	6.0465	2.1263	61.429
3500	1550	1000	0.5792	0.9693	0.3618	6.0194	2.2095	107.82
3500	1600	1000	0.5708	0.9574	0.3462	5.9928	2.2935	149.94
3500	1650	1000	0.5621	0.9465	0.3307	5.9664	2.3786	188.58
3500	*1700*	*1000*	*0.5532*	*0.9366*	*0.3152*	*5.9401*	*2.4651*	*224.32*
3500	*1750*	*1000*	*0.5440*	*0.9273*	*0.2997*	*5.9137*	*2.5535*	*257.63*

Note: The following assumptions are made: (a) The density of the satellite's central portion ρ_c is taken constant at a value corresponding to the mean density of Io to allow for the presence of a rock-metal component. (b) The density of the water-ice/liquid outer shell ρ_s is representative for that of clean ice I separated from a mixture of high-pressure water ice and rock-metal component underneath. (c) The possible range of meaningful models is limited by the pressure-induced transition from ice I to ice II at around 209 MPa. Emphasized in italic are model determinations that supersede that critical pressure at the base of the outer ice shell.

Neptune's largest moon Triton during its 40,000 km approach to that satellite. Neptune and Triton are the most distant planetary objects, yet visited by spacecraft. Data on the masses, radii, and

Table 23. Mean radius R, mass GM, mean density $\bar{\rho}$, and axes of the largest satellites of Uranus and Neptune's largest moon *Triton*. a, b, c denote the planet-facing, orbit-facing, and polar axis, respectively. References: [07Sei, 91Jac, 92Jac].

satellite	R [km]	GM [km^{3}s^{-2}]	$\bar{\rho}$ [kg m^{-3}]	a [km]	b [km]	c [km]
Miranda	235.8 ± 0.7	4.4 ± 0.5	1200 ± 140	240.4 ± 0.6	234.2 ± 0.9	232.9 ± 1.2
Ariel	578.9 ± 0.6	90.3 ± 8	1660 ± 150	581.1 ± 0.9	577.9 ± 0.6	577.7 ± 1.0
Umbriel	584.7 ± 2.8	78.2 ± 9	1390 ± 160			
Titania	788.9 ± 1.8	235.3 ± 6	1710 ± 50			
Oberon	761.4 ± 2.6	201.1 ± 5	1630 ± 50			
Triton	1352.6 ± 2.4	1427.9 ± 3.5	2061 ± 7			

densities of the Uranian satellites are given in Table 23. Only the inner satellites Miranda and Ariel have significant variations in the main axes and are best described by triaxial ellipsoids. Because most of the data was obtained during one flyby of *Voyager 2* through the Uranus system, the uncertainties in the values for density, mass and shapes of the Uranian satellites are significantly greater as compared to the data for the Jupiter and Saturn systems.

We refer the reader to [07Hus] for a more complete account on the interiors of the Uranian and Neptunian satellites.

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