10.01 Physics of Terrestrial Planets and Moons: An Introduction and Overview

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10.01.1 Introduction

Humanity has always been fascinated with the wandering stars in the sky, the planets. Ancient astrologists have observed and used the paths of the planets in the sky to time the seasons and to predict the future. Observations of the planets helped J. Kepler to formulate his laws of planetary motion and revolutionize the perception of the world. With the advent of the space age, the planets have been transferred from bright spots in the sky to worlds of their own right that can be explored, in part by using the in situ and remote sensing tools of the geosciences. The terrestrial planets are of particular interest to the geoscientists because comparison with our own planet allows a better understanding of our home, the Earth. Venus offers an example of a runaway greenhouse that has resulted in what we would call a hellish place. With temperatures of around 450 °C and a corrosive atmosphere that is also optically nontransparent, Venus poses enormous difficulties to spacecraft exploration. Mars is a much friendlier planet to explore but a planet where greenhouse effects and atmospheric loss processes have resulted in a cold and dusty desert. But aside from considerations of the usefulness of space exploration in terms of understanding the Earth, the interested mind can visit astounding and puzzling places. There is the dynamic atmosphere of Jupiter with a giant thunderstorm that has been raging for centuries. There is Saturn with its majestic rings, and there are Uranus and Neptune with complicated magnetic fields. These giant planets have moons that are astounding. There is the volcanic satellite of Jupiter, Io, that surpasses the Earth and any other terrestrial planet in volcanic activity. This activity is powered by tides that twist the planet such that its interior partially melts. A much smaller moon of Saturn, Enceladus, also has geysers that could be powered by tidal heating although the case is less clear. Its volcanic activity releases water vapor not lava, however. There is another moon of Saturn, Titan, that hides its surface underneath a layer of photochemical smog in a thick nitrogen atmosphere, and there are moons of similar sizes that lack any comparable atmosphere. Miranda, a satellite of Uranus, appears as if it has been ripped apart and reassembled. And Triton, the major satellite of Neptune, has geysers of nitrogen powered by solar irradiation. Magnetic field data suggest that icy moons orbiting the giant planets may have oceans underneath thick ice covers. These oceans can, at least in principle, harbor or have harbored life. Moreover, there are asteroids with moons and comets that may still harbor the clues to how the solar system and life on Earth formed.

This volume of the Treatise on Geophysics discusses fundamental aspects of the science of the planets. It is focused on geophysical properties of the Earth-like planets and moons, those bodies that consist largely of rock, iron, and water, and the processes occurring in their interiors and on their surfaces. But it goes further by discussing the giant planets and asteroids and comets as well and other extrasolar planetary systems. The better part of the volume is dedicated to the formation, interior structure, and evolution of the terrestrial planets and to their physical properties such as gravity and magnetic fields, rotation, the physics of the atmosphere, and surface–atmosphere interactions. What is the planetological context of life and are there feedbacks between life and planetary evolution? An attempt at answering these questions is the subject of a dedicated chapter. In the second part, we turn to the outer fringes of the solar system beyond the asteroid belt, to the giant planets and their satellites, to Pluto and the Kuiper Belt, and, finally, to extrasolar planets. Because exploration by spacecraft is so fundamental to the geophysics of the planets, the volume has two chapters that discuss the orbital dynamics – the routes to the planets – and the instruments typically flown on planetary missions. This edition of the volume has been widened in scope and coverage in comparison with the previous editions with dedicated chapters on asteroids and comets, on planetary tectonics and volcanism, on planetary atmospheres, and on extrasolar planetary systems.

10.01.2 Our Planetary System

The solar system contains a myriad of bodies ranging in size from the Sun to miniscule dust particles. The *Encyclopedia of Planetary Sciences* (Shirley and Fairbridge, 1997; on the web as part of the *Earth Sciences Series* at www.springerreference.com), *Encyclopedia*

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of Astrobiology (Gargaud et al., 2011), and *Encyclopedia of the Solar System* (Spohn et al., 2014); Vol. VI-4-B of the Landolt–Börnstein Series (Trümper, 2009); the Planetary Scientist's Companion (Lodders and Fegley, 1998); and textbooks such as de Pater and Lissauer (2010) are useful sources of information about the solar system. Valuable collections of planetary data can be found on http://science.jpl.nasa.gov/PlanetaryScience, on http://ssd.jpl.nasa.gov, and on http://nssdc.gsfc.nasa.gov. A collection of images of planetary surfaces can be found in the NASA Photojournal (http://photojournal.jpl.nasa.gov).

The eight planets Mercury, Venus, Earth, Mars, Jupiter, Saturn, Uranus, and Neptune are shown in Figure 1 together with Pluto, the most prominent member of a new class of objects, the dwarf planets. The International Astronomical Union (IAU) introduced the class of dwarf planets in 2006. These are intermediate in size between the terrestrial planets and the 'small solar system bodies' such as cometary nuclei and most asteroids. The IAU has so far identified the following five celestial bodies as dwarf planets – Pluto, the asteroid Ceres, and the trans-Neptunian objects (TNO) Eris, Haumea, and Makemake – but there should be many more still to be recognized dwarf planets in the solar system (e.g., http://en.wikipedia.org/wiki/List_of_possible_dwarf_planets). A planet of our solar system according to the IAU resolution 5 (http://www.iau.org/news/pressreleases/detail/iau0603/#1) is a celestial body that (a) is in orbit around the Sun, (b) has sufficient mass for its self-gravity to overcome rigid body forces so that it assumes a hydrostatic equilibrium (nearly round) shape, and (c) has cleared the neighborhood around its orbit. A dwarf planet according to the resolution satisfies (a) and (b) but has *not* cleared its neighborhood and (d) is not a satellite. All other objects, except satellites – so the resolution continues – shall be referred to collectively as 'small solar system bodies.' See also Basri and Brown (2006) for a discussion of the term 'planet.'

The Sun, a middle-aged main sequence star, contains 98.8% of the mass of the solar system but only 0.5% of its angular momentum. The next smaller body, Jupiter, still 300 times more massive than the Earth (see **Table 1**), contains more than 60% of the mass of the rest. Jupiter is the biggest of the *giant planets*, a group of gaseous planets that constitute a major subgroup of the solar system and other known planetary systems (see **Chapters 10.16** and **10.21**). Among the giant planets are, in addition to Jupiter, Saturn, Uranus, and Neptune. The latter two are sometimes called the subgiants or the ice giants because they are notably smaller than Saturn and Jupiter and because they mostly consist of water, methane, and ammonia, components often collectively called the planetary ices (see below). The Earth, the biggest member of the other major subgroup, the *terrestrial planets*, is the only planet on which we know to date that life has originated and evolved. Among the members of this group are Mercury, the innermost planet; Venus, the Earth's twin with respect to size and mass; and Mars. The latter planet has the best chance of having – or having once had – some primitive forms of life, which makes it a prime target of present-day space missions. The terrestrial planets together have about 0.005% of the mass of the solar system. **Table 1** collects some data of general interest on the planets, dwarf planets, and some major moons.

Comparative planetology is the science of studying the planets by comparing and finding general properties and common lines of evolution as well as features that are specific. It uses the methods of the natural sciences and is strongly interdisciplinary. Comparative planetology is not restricted to the planets *sensu stricto* but also considers the major moons of the planets such as the Earth's Moon, the major satellites of Jupiter (Io, Europa, Ganymede, and Callisto; **Figure 2**), the major Saturnian satellite Titan, and, finally, Triton, Neptune's major satellite. Other bodies of interest include the yet largely unexplored members of the asteroid belt, the asteroids, the comets, and the Kuiper Belt

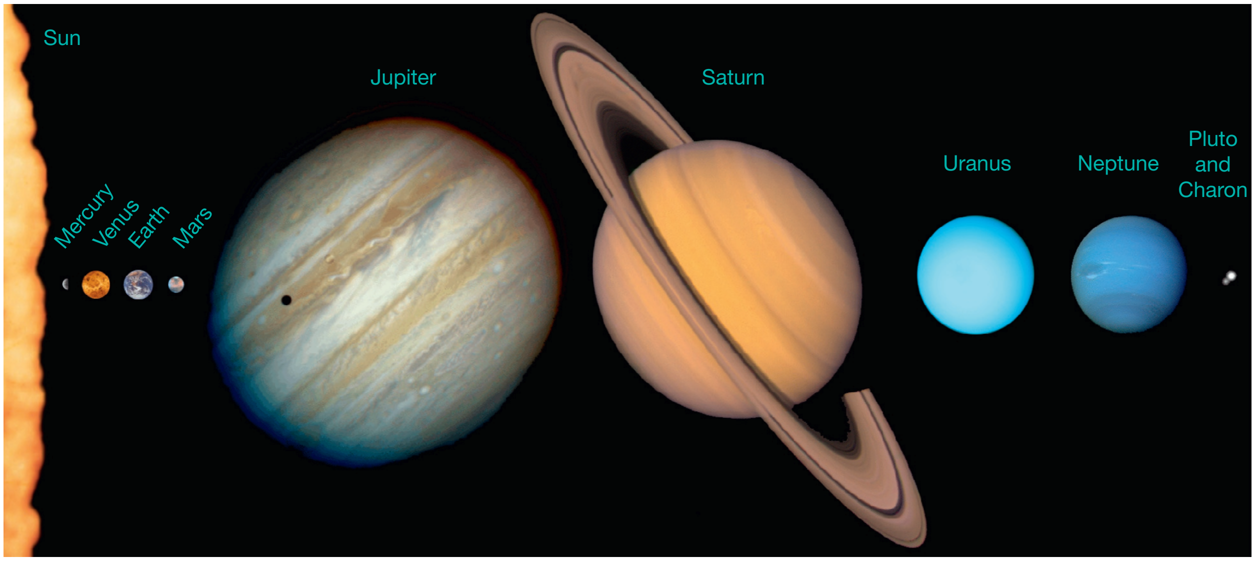

Figure 1 The eight planets Mercury, Venus, Earth, Mars, Jupiter, Saturn, Uranus, and Neptune and the dwarf planet Pluto are shown in this compilation of NASA images with their correct relative sizes and ordered according to their distance from the Sun. Mercury is barely visible at left close to the arc of the surface of the Sun, and Pluto is barely visible at the outer right. The dark spot on Jupiter is the shadow of Io, one of its major satellites. © C. J. Hamilton.

Table 1 Some data of general interest on the planets, dwarf planets, and some major moons

(a) Properties of the planets

	Terrestrial or Earth-like planets				*Giant planets*			
	Mercury	*Venus*	*Earth*	*Mars*	*Jupiter*	*Saturn*	*Uranus*	*Neptune*
Radius (km)	2438.0	6052.0	6371.0	3390.0	71492.0	60268.0	24973.0	24764.0
Mass (10^{24} kg)	0.3301	4.869	5.974	0.6419	1899.0	568.46	86.83	102.4
Density (10^3 kg m^{-3})	5.430	5.243	5.515	3.934	1.326	0. 6873	1.318	1.638
Uncompressed density (kg m^{-3})	5.3	4.0	4.05	3.75	0.1	0.1	0.3	0.3
Rotational period (d[a])	58.65	243.0[b]	0.9973	1.026	0.4135	0.4440	0.7183[b]	0.6713
Inclination of rotation axis (°)	0.03	177.3	23.44	25.19	3.13	26.73	97.8	28.32
Orbital distance (AU[c])	0.3871	0.7233	1.000	1.524	5.203	9.572	19.19	30.07
Orbital period (a[d])	0.2410	0.6156	1.001	1.882	11.87	29.39	84.16	165.0
Magnetic moment (10^{-4} T $\times$ radius^3)	3×10^{-3}	$<3\times10^{-4}$	0.61	$<6\times10^{-4}$	4.3	0.21	0.23	0.133
Effective surface temperature (K)	445	325	277	225	123	90	63	50
Specific heat flow or luminosity (pW kg^{-1})	?	?	7	?	176	152	4	67
Known satellites	0	0	1	2	65	62	27	13

(b) Properties of dwarf planets (ordered by their distance from the Sun)

	Ceres	*Pluto*	*Haumea*	*Makemake*	*Eris*
Radius (km)	476.2	1151.0	620	715	1200 ± 50
Mass (10^{24} kg)	9.47×10^{-4}	0.01309	?	?	?
Density (10^3 kg m^{-3})	2.09	2.050	?	?	?
Rotational period (d[a])	0.378	−6.387[b]	0.163	?	?
Orbital distance (semi major axis) (AU[c])	2.765	39.48	43.00	45.34	67.73
Orbital period (a[d])	4.6033	248.09	282,12	305,55	561.74
Orbital eccentricity	0.07913	0.249	0.19813	0.16384	0.4336
Orbital inclination (°)	10.59	17.14	28.22	29.00	46.87
Effective surface temperature (K)	167	44	34	36	~30
Known satellites	0	5 (Charon, Nix, Hydra, Kerberos, Styx)	2 (Namaka, Hi'iaka)	0	1 (Dysnomia)

(c) Properties of major satellites

	Moon	*Io*	*Europa*	*Ganymede*	*Callisto*	*Titan*	*Triton*
Primary	Earth	Jupiter				Saturn	Neptune
Radius (km)	1737.4	1821.5	1562.1	2632.3	2409.3	2575.5	1352.6
Mass (10^{20} kg)	734.9	893.2	480.0	1481.9	1075.9	1346.0	214.0
Density (10^3 kg m^{-3})	3.344	3.53	3.02	1.94	1.85	1.881	2.061
Specific heat flow (pW kg^{-1})	8	890	150(?)	?	?	?	?
Orbital period (d)	27.32	1.769	3.551	7.155	16.69	15.95	5.877[b]

For the sake of space, we give four significant digits even for numbers that are more accurately known.
The uncompressed density is model dependent and cannot be given with more than 2–3 significant digits.
[a]A day (d) is equivalent to 24 h.
[b]The motion (rotation, revolution) is retrograde.
[c]AU is one astronomical unit or 149.6 million kilometers.
[d]A year (a) is equivalent to 365 days (d).
Source: The data have been taken from the compilation of Lodders and Fegley, The Planetary Companion, Oxford University Press, 1998, from Spohn, Planetologie, in Bergmann-Schäfer, Lehrbuch der Experimentalphysik, Vol. 7, (ed. W. Raith), W. De Gruyter, Berlin, p. 427–525 and from http://nssdc.gsfc.nasa.gov and http://ssd.nasa.gov.

objects. The asteroid belt is found between Mars and Jupiter, and its members are generally accepted to be the parent bodies of most meteorites (stones from space found on the Earth's surface; see Chapter 10.15). The Kuiper Belt stretches between 30 AU (roughly the orbit of Neptune) and at least 50 AU and contains the dwarf planets Pluto, Haumea, Makemake, and Eris (Chapters 10.19 and 10.20). The asteroids are believed to be the source of most meteorites and contain rich information about the formation of the solar system and its evolution and even the evolution of matter before the formation of the solar system. The Kuiper Belt objects are largely unexplored. Pluto, the most prominent member, and the Kuiper Belt are the target of the New Horizons mission launched in 2006 and scheduled to arrive at Pluto in 2015 (see Chapters 10.19 and 10.20). The emerging field

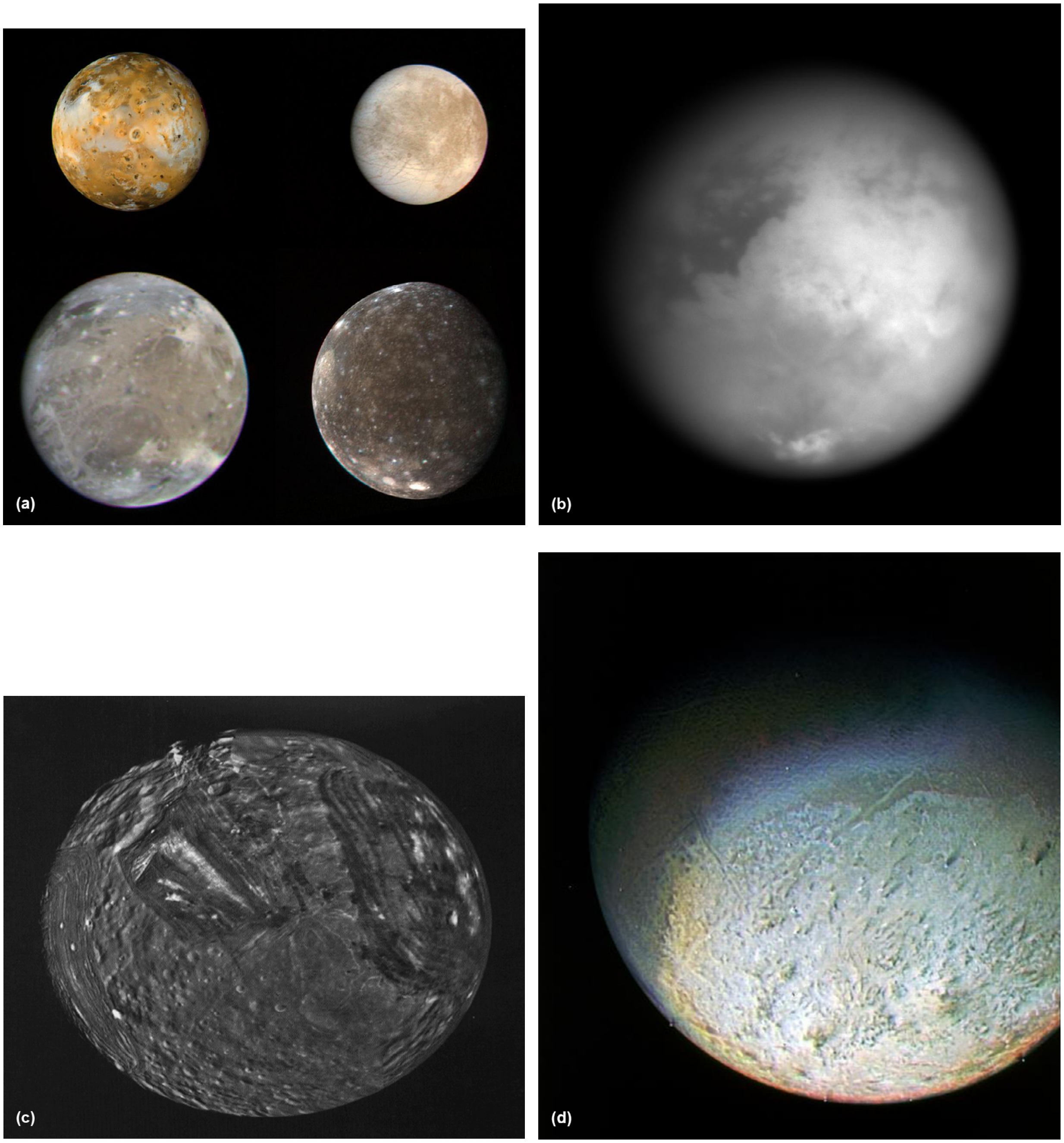

Figure 2 The four Galilean satellites of Jupiter Io, Europa, Ganymede, and Callisto (a), the Saturnian satellite Titan (b), the Uranian satellite Miranda (c), and the Neptunian satellite Triton (d) are shown in this compilation of spacecraft images from top left to bottom right. NASA/JPL.

of extrasolar planetology will further widen the scope of comparative planetology.

10.01.3 Planetary Missions

The results of planetary sciences have been made possible at their present levels only through the exploration of the solar system with space missions (see, e.g., Burke, 2014 and Chapters 10.22 and 10.23. See also a list of missions to the planets in Spohn et al. (2014)). All planets and major satellites of the solar system have been visited at least by a flyby of a robot spacecraft. This is not true of the dwarf planets, but missions are on their ways to Pluto (New Horizons) and Ceres (Dawn). Dawn has orbited Vesta between July 2011 and September 2012 (Russel and Raymond, 2011 and Chapter 10.15) and is planned to go into orbit around Ceres in February 2015. New Horizons is planned to flyby Pluto and its satellites with closest approach in June 2015 (Chapters 10.19 and 10.20). Some, such as Venus, the Moon, Mars, Jupiter, and Saturn, have been explored by orbiters and by landers. The Moon has even been visited by astronauts. It can be said without any doubt that space exploration has turned dim disks in the sky observable with telescopes into worlds of their own rights waiting for further, perhaps eventually, human exploration.

Space exploration began with missions to the Moon in 1959, characterized by the race to the Moon between the United States and the then Soviet Union. This race culminated in the landing of men on the Moon. For planetary sciences, the significance of the Apollo program lies mainly with the first sample return, the first seismic exploration of a planetary interior other than the Earth's, the first heat flow measurement, and the first geologic field trip on another planet. Although Apollo has not finally solved the puzzle of the origin of the Moon – some may argue that it can never be solved once and for all – it has provided a wealth of data and has made the Moon one of the best-known planetary bodies. Scientists continue to successfully explore the Apollo data as the recent discovery of water in lunar rock (Saal et al., 2008) and the seismic detection of the core (Weber et al., 2011) have demonstrated. Missions to the Moon have been accomplished by Japan, China, India, and ESA (Kaguya, Chang'e, Chandrayaan, and Smart-1). A recent highlight of geophysical exploration of the Moon has been the GRAIL (Gravity Recovery and Interior Laboratory) mission of NASA with two spacecraft doing gravity gradiometry (**Chapter 10.05**).

Highlights of Mars exploration include the Viking missions, the first landings on Mars; the Pathfinder mission, the first vehicle introducing mobility; Mars Express, the first European planetary mission *sensu stricto*; the Mars Exploration Rovers, MER for short; and the Mars Science Laboratory on the Curiosity rover. From a geophysics point of view, Mars Global Surveyor, which returned highly accurate maps of topography and gravity (Smith et al., 1999) and detected the remnant magnetization of old crustal units (Acuña et al., 1999), has been a particularly remarkable mission. The MER (**Figure 3**) have been doing the first geologic field trips on Mars for the past 10 years and detected on their now tens of kilometers long journey minerals that formed under humid conditions (Squyres and Knoll, 2005 and papers therein). The Mars Express mission detected methane on Mars, numerous morphological units that required water to form (**Figure 4**, **Chapter 10.11**) and showed that volcanic activity continued albeit at a low rate until the recent past (Neukum et al., 2004). Mars Science Laboratory/Curiosity, more massive than any previous planetary exploration rover and equipped with a most sophisticated payload, is exploring the Gale crater and has found evidence for a past environment well suited to supporting microbial life. The first *in situ* geophysical observatory InSight equipped with a very broad band seismometer (compare **Chapter 10.03**), a heat flow probe, and a magnetometer is scheduled for launch in 2016. The lander will use its communication hardware for radio science tracking of, for example, the rotation axis of the planet. A sample return mission to Mars continues to be high on the agenda of the planetary science community.

An important driver for Mars research is its potential habitability and the possible detection of life, extinct or extant. The Viking landers in the 1970s already attempted to find signatures of bioactivity but for most failed to do so. Beagle 2 onboard the European Mars Express Mission was built to search for bioactivity and would have been the first European lander on another planet had it not failed. The next attempt to directly detect biosignatures on Mars will be the ESA ExoMars rover rescheduled for launch in 2018.

Venus has been the target of very successful Soviet missions that included eight Landers (Venera 7–14). These landers survived in the highly corrosive atmosphere of Venus only for a

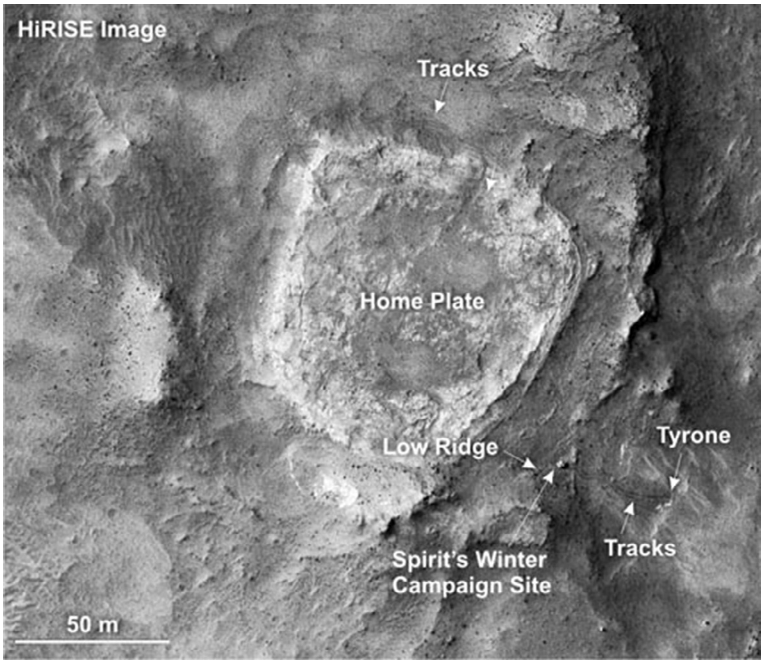

Figure 3 Traces of the MER rovers imaged by the ultimate-resolution camera onboard the Mars Reconnaissance Orbiter. NASA/JPL.

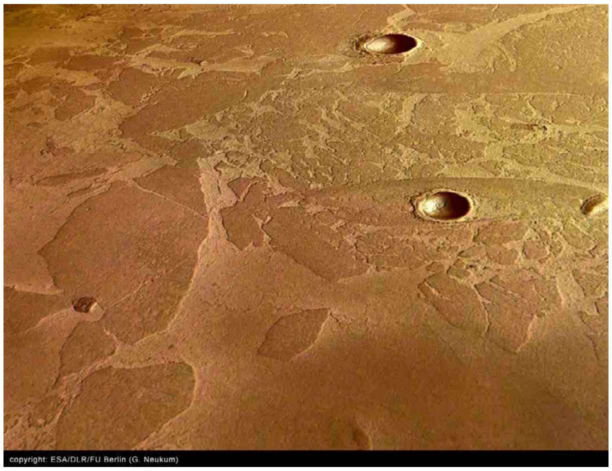

Figure 4 This image has been interpreted to show a frozen over lake on Mars comparable in size to the North Sea on Earth. © ESA/DLR/FU Berlin (G. Neukum).

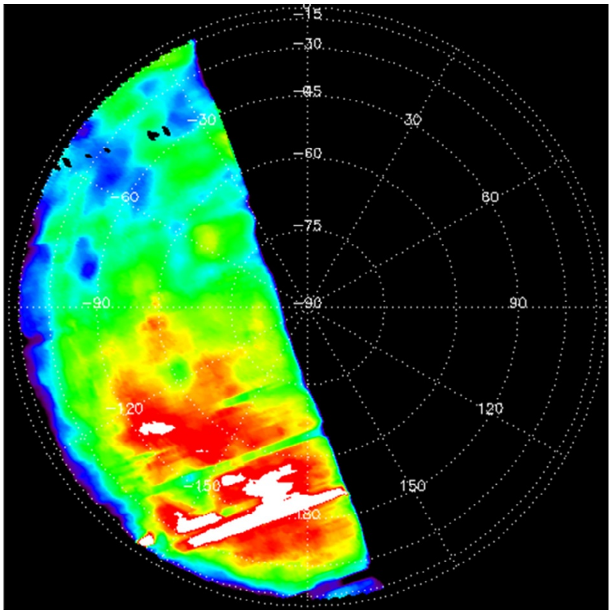

Figure 5 Map of half of the southern hemisphere of Venus in the infrared taken with the VIRTIS infrared spectrometer on Venus Express. © ESA.

few hours but transmitted color photos from the surface that shape our image of the Venusian surface. Further highlights of Venus exploration have been the Pioneer Venus, Magellan, and Venus Express missions, which provided extensive data on the atmosphere of Venus and the first systematic mapping of the surface with radar and in the infrared (Figure 5). Since the atmosphere of Venus is optically thick, cameras are of little use at this planet for surface exploration but are used to study atmosphere circulation patterns.

Mercury, the innermost planet, was visited by two flybys of the American Mariner 10 mission targeted for Mars in the 1970s. But since March 2011, the NASA MESSENGER mission has been orbiting the planet and has provided a wealth of data. The next mission to the planet will be the ambitious European BepiColombo mission scheduled for launch in 2016 with orbit insertion in 2024 and with a planetary remote sensing orbiter and a second orbiter – provided by the Japanese space agency JAXA – to explore the Hermean magnetosphere.

The outer solar system was first explored by the Pioneers that flew by Jupiter (Pioneers 10 and 11) and Saturn (Pioneer 11) and subsequently by the very successful Voyager I and II missions. The latter flew by all four major planets of the outer solar system. In addition to exploring the planets, these spacecraft discovered rings at Jupiter and Uranus and explored the satellite systems and found Io to be the most volcanically active body in the solar system. The Voyagers for the first time allowed a detailed view of the rich phenomenology of icy satellite surfaces. After the Pioneer and Voyager flybys, orbiters were the next logical step in outer solar system exploration. The first mission in this class was Galileo that orbited Jupiter and plunged into Jupiter's crushing atmosphere in September 2003. The spacecraft was deliberately destroyed to protect the Jovian system, in particular Europa, from being polluted. Galileo discovered the magnetic field of Ganymede and evidence for oceans underneath the ice crusts of the Galilean moons Europa, Ganymede, and Callisto. On its route to Jupiter, Galileo discovered for the first time a moon (Dactyl) that is orbiting an asteroid (Ida) and observed comet Shoemaker–Levy crashing into Jupiter. Galileo imaged the Jovian satellites at much improved resolution and discovered several new ones. Comparison with Voyager images of volcanic features on Io showed significant modifications that had occurred in the roughly 20 years between the two missions. An instrumented descent probe released from the Galileo orbiter entered the Jovian atmosphere in December 1995 and provided the first *in situ* measurements of the state and the chemistry of a giant planet atmosphere shroud. The results of the Galileo mission are discussed in Bagenal et al. (2004). The NASA Juno mission scheduled to go into a polar orbit in July 2016 to study the planet's gravity and magnetic fields, its atmosphere dynamics, and its composition will revisit Jupiter. The Jovian satellites will be the prime target of ESA's JUICE (JUpiter ICy moons Explorer) mission in the 2020s. Among the most spectacular planned missions to the Jovian system is a lander mission for Europa either as a Russian contribution to the JUICE mission or as a NASA mission. The mission plans are driven by the perception that an ocean on Europa may be covered by a thin ice shell that can be drilled and by the possibility that life developed in this ocean. It is possible that water from this ocean may have surfaced through cracks in the ice shell and may carry chemical signals of a biosphere.

At the time of this writing, Cassini has been orbiting Saturn for a decade. Cassini has even topped the success of Galileo as an outer solar system exploration mission through its more advanced payload capabilities and longer mission duration. The Huygens probe supplied by the European Space Agency (ESA) descended through the atmosphere of Saturn's biggest moon Titan and successfully landed on the surface on 14 January 2005 to transmit the first images (color) from Titan's surface (Figure 6). Titan's atmosphere is optically thick like that of Venus. Huygens landed on a solid surface strewn with ice boulders. On its descent, it imaged rivers in which

Figure 6 First color image from Titan's surface. This image was processed from images returned on 14 January 2005, by ESA's Huygens probe during its successful descent to land on Titan. The rocklike objects just below the middle of the image are about 10 cm across. The surface consists of a mixture of water and hydrocarbon ice. There is also evidence of erosion at the base of these objects, indicating possible fluvial activity. © ESA/NASA/JPL/ University of Arizona.

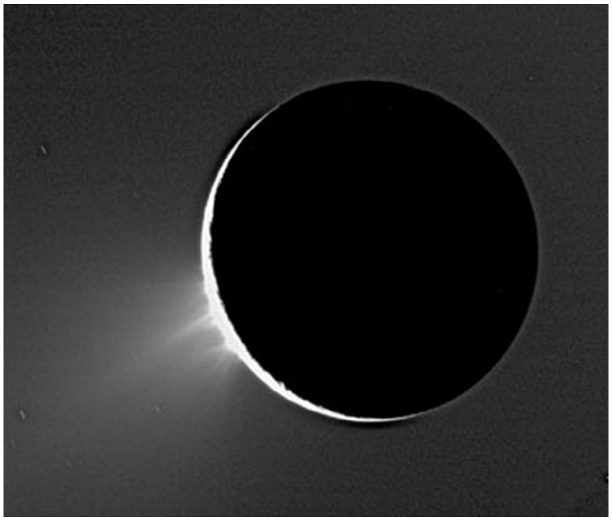

Figure 7 Geysers emanating from the southern hemisphere of Enceladus. NASA/JPL.

hydrocarbons are flowing or have flowed in the past and a possible shoreline (Lebreton et al., 2005). Near-infrared data from the Cassini mission have been interpreted as showing cryovolcanoes on Titan (e.g., Sotin et al., 2005). Another highlight of the mission is the detection of active geysers on the small satellite Enceladus (Porco et al., 2006, **Figure 7**), the diameter of which is only 500 km, and a rift structure circling Iapetus along its equator (Porco et al., 2005). Cassini also returned gravity data that allowed models of the interiors of some of the smaller satellites (see **Chapter 10.18**). Most recently, Cassini gravity data taken at Enceladus have been interpreted to support the existence of the long speculated (partial) ocean that could feed the geysers (Iess et al., 2014). This ocean would not be global but extend underneath most of the southerly hemisphere and has been speculated to be another potential habitat. Cassini will continue to explore the Saturn system through the Cassini Solstice Mission until September 2017 and then plunge into Saturn. Mission ideas to follow in the footsteps of Cassini include dedicated Titan and Enceladus Explorers. Just as for the Europa missions, these proposal are at least in part driven by the interest in possible extraterrestrial life.

Asteroids and comets are believed to be remnants of accretion and to consist of largely primitive matter (see **Chapter 10.15**). Although this perception is questionable to some extent – comet nuclei that have repeatedly flown by the Sun may have experienced thermal metamorphism; some asteroids show signs of endogenic activity and may be differentiated – asteroids and comets are minor solar system bodies of great interest. The Dawn mission has explored the second largest asteroid Vesta and is on its way to the dwarf planet Ceres. Vesta has been shown to be a differentiated body, while the latter could be undifferentiated and mostly primitive. Comet nucleus Churyumov–Gerasimenko is the target of the ESA Rosetta mission launched in 2004. After a 10-year cruise, the Rosetta orbiter and the lander Philae have ended hibernation and will begin to orbit the nucleus in late summer of 2014, while the lander Philae will dock onto the nucleus in the fall and explore it *in situ*. (We are speaking of docking rather than landing because of the small gravity of the nucleus.) Both the orbiter and Philae will take data as the comet approaches the Sun.

Planetary exploration has taken and is still following a stepped approach by which flybys are followed by orbiter missions. These are followed by lander missions including networks of landing stations and rovers. Sample return to the Earth is then foreseen, followed by human exploration. The latter step is often met with skepticism because of its prohibitive cost and is still facing huge problems of technical and medical nature and of safety for the astronauts.

The payloads of orbiters and flyby spacecraft are specifically designed for their science goals (see **Chapter 10.23**). It is still possible, however, to characterize generic payload elements that are typically found on these spacecraft. Almost every mission has at least one camera, often two for wide angle and for high resolution. Early missions had panchromatic cameras, but recent missions have color capabilities. Stereo is a useful feature that can increasingly be found on missions like Mars Express. Stereo allows 3-D imaging and enables the development of digital terrain models. Digital terrain modeling is helped by laser altimetry as the successful Mars Observer Laser Altimeter (MOLA) on Mars Global Surveyor has so convincingly demonstrated. If the surface is hidden underneath an optically thick atmosphere, radar can be used to map the surface as has been successfully done at Venus by the Magellan mission. Radar is further used to map the surface roughness and to penetrate into the first centimeters of the near-surface layers, for instance, to look for ground ice. Other microwave sounders have been proposed that use different wavelengths and have differing depths of penetration.

While cameras, laser altimeters (**Figure 8**), and radio science instrumentation are indispensable instruments for planetary geology, geophysics, and geodesy, spectrometers are the tools of cosmochemistry and mineralogy. These instruments typically analyze particles (photons, neutrals, and ions) emitted from surfaces and atmospheres. Visible, near-infrared, and thermal infrared (0.5–15 μm) spectrometers are used to characterize the mineralogy of surface rock (silicates, carbonates, sulfates, etc.) and to search for water ice. Mapping infrared spectrometers such as Themis on Mars Odyssey and Omega on Mars Express allow a mapping of minerals on the surface. The Planetary Fourier Spectrometer on Mars Express works in a similar but broader spectral range (1.2–45 μm) but has a much higher spectral resolution at the expense of spatial resolution. Ultraviolet spectrometers measure neutral metals such as Al, Na, and S, ozone, and OH radicals. Gamma-ray spectrometers can measure radio nuclides (K, Th, and U), major rock-forming elements (O, Mg, Al, Si, Ca, Ti, and Fe), hydrogen (from water), and carbon, depending on the energy resolution and the elemental concentration in the surface. x-Ray spectrometers are suitable to detect the major elements Mg, Al, Si, Ca, and Fe.

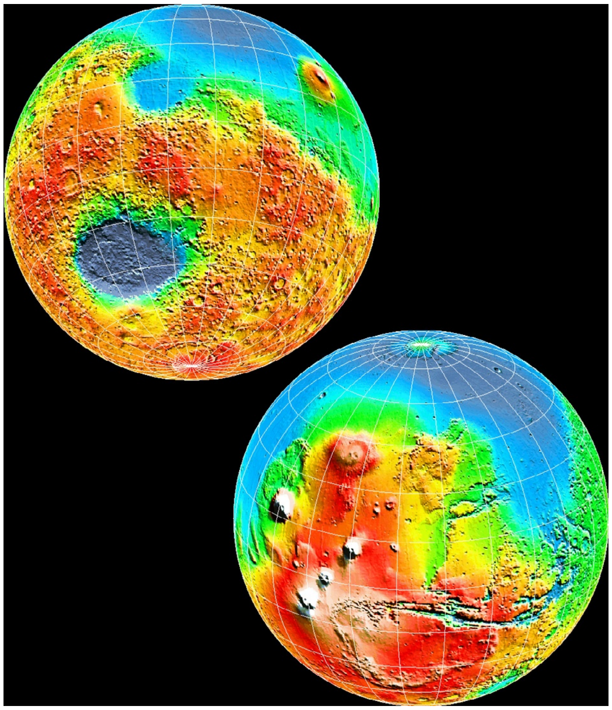

Figure 8 Global map of Mars taken by the Mars Observer Laser Altimeter (MOLA) onboard the Mars Global Surveyor mission. NASA/JPL.

The particles and electromagnetic fields surrounding planets and other solar system objects are measured with magnetometers and plasma detectors. Dust analyzers, finally, collect dust particles and analyze the direction and strength of dust flux in space.

Landing missions carry cameras and have been equipped with spectrometers and other sensors to measure composition and mineralogy. Mobility is an important issue since exploration of just one landing side may lead to biased results. Recent examples are the two MER rovers that carry a panoramic stereo camera, an alpha-particle x-ray spectrometer for elemental composition, a Mössbauer spectrometer for the identification of iron-bearing minerals, a miniaturized thermal infrared spectrometer for mineralogy, and a rock-abrasion tool to allow the measurement of 'fresh' rock (Squyres et al., 2003). Curiosity has a laser-induced breakdown spectrometer, an x-ray diffraction and fluorescence instrument, a neutron spectrometer, an environmental monitor package, a hand-lens imager, and a radiation assessment detector (Grotzinger et al., 2012). Network landers have not been flown yet but are generally agreed to be important (compare also **Chapter 10.03**). These landers will operate the same payload simultaneously at several locations. Applications are seismology to explore the interiors of planets, meteorology to explore planetary atmospheres, electromagnetic induction studies, and planetary geodesy. The NASA InSight mission will place a geophysical station on the surface of Mars in 2016. ESA is preparing for the ExoMars mission with the Pasteur rover to be launched in 2018 that will search for traces of extinct and extant life.

The search for extrasolar planets mostly uses two technologies, transit observations and radial velocity measurements through Doppler shift observations (see **Chapter 10.21**). While the former uses the slight dimming of the light of the central star as the planet passes between the star and the observer, the latter uses the Doppler shift of light from the star induced by the gravity pull of the planet on the star. Radial velocity measurements are done with ground-based telescopes or interferometers (such as the HARPS near Geneva, Switzerland, http://www.eso.org/sci/facilities/lasilla/instruments/harps.html) and constrain the mass of the planet. Transits are observed using ground-based and space telescopes such as COROT and Kepler and measure the radius of the planet. A planet observed with both methods can be characterized by an average density. Newly detected planets are candidates until they are confirmed by backup observation. Several thousand candidates have been identified (most of them by Kepler), and roughly two thousand are confirmed. Future missions (such as PLATO (ESA) and TESS (NASA)) aim to detect planets of Earth size and smaller size orbiting bright stars and to do spectroscopy.

10.01.4 Planet and Satellite Orbits and Rotation States

There are some interesting commonalities between the orbits and rotation states of the planets and their major satellites. The rotation of the terrestrial planets is discussed in detail in

Chapter 10.04. The orbits lie mostly in a plane that is defined by the Earth's orbital plane, termed the ecliptic, and that is close to the equatorial plane of the Sun. The formal definition of the ecliptic plane adopted by the IAU calls for the ecliptic pole to be the mean orbital angular momentum vector of the Earth–Moon barycenter.

The normals to the actual orbital planes have inclinations relative to the ecliptic normal that are small, only a few degrees. The dwarf planets differ with orbital inclinations between 11° and 47° (Table 1). Small bodies (asteroids and comets) also tend to have larger orbital inclinations. The rotation axes of most planets and major satellites are within a few tens of degrees to the vertical to their orbital planes. Notable exceptions are Uranus, whose rotation axis lies almost in its orbital plane; Venus, whose retrograde rotation can be expressed as an inclination of almost 180°; and Pluto, whose inclination is between those of Uranus and Venus. The reasons for these anomalies are unknown but are sometimes speculated to be attributable to impacts of planet-size bodies on the young planets during the late stage of accretion.

The rotation and revolution of most planets and moons are prograde, that is, counterclockwise if viewed from above the celestial north pole. Exceptions are Venus, Uranus, and Pluto whose retrograde rotations can also be described as inclinations of more than 90° of their rotation axes to their orbital plane normals. Another exception is Triton, whose retrograde orbital motion about Neptune is unusual. Triton, like Pluto, is speculated to have originated in the Kuiper Belt. It was likely captured into its retrograde orbit by Neptune (see Chapter 10.17). There are many more satellites with retrograde orbits. However, these irregular satellites are typically much smaller than Triton, objects in a size range of only a few kilometers to a few tens of kilometers. The known rotation periods of the major satellites and of most satellites in general are equal to their orbital periods. This is termed a 1:1 spin–orbit coupling and is believed to be the result of tidal evolution (see Chapters 10.17 and 10.18). Tidal evolution occurs because planets raise tides on their satellites just as the Moon raises tides on the Earth. The gravitational force of the planet then pulls on the tidal bulge of the satellite. This torque will reduce the rotation rate of the satellite – should it be greater than its average orbital angular speed – and the interaction will end with the observed 1:1 coupling. As a result, the satellites present their primaries with mostly the same face at all times.

The planet Mercury is particularly interesting with respect to spin–orbit coupling since it is the only planet that is in such a resonance state with the Sun. However, its coupling is not 1:1, but 3:2 (see Table 1). The reason for this somewhat odd ratio is agreed to lie with the unusually large eccentricity of Mercury's orbit that has a value of 0.2 (Colombo, 1965; Chapter 10.04). Values of eccentricity <0.1 are more typical of planets and major satellites; the dwarf planets and small bodies have larger eccentricities. The large eccentricity of Mercury's orbit causes significant differences to arise between the constant rotational angular velocity and the orbital angular velocity that varies along the orbit. The two angular velocities could only be exactly equal at all times if the orbit were perfectly circular with zero eccentricity. The orbital velocity on an eccentric orbit increases towards the perihelion (the point of the orbit closest to the Sun) and decreases towards the aphelion (the point farthest from the Sun). Mercury's 3:2 resonance causes the orbital angular velocity to be equal to the rotational angular velocity at perihelion. This minimizes the tidal torque on Mercury and stabilizes the resonance.

The orbital distances of the planets from the Sun roughly follow a law with the distance of one planet to the Sun being roughly twice the distance to its inner neighbor. This rule is called the *Titius–Bode law* and works with Jupiter and Mars only if the asteroid belt is counted as a planet. The origin of the law is little understood. Regular relations between orbital distances or periods (the latter two are coupled through Kepler's third law of orbital motion) are not unusual in the solar system, however. The most prominent example is the Laplace resonance between the innermost three major satellites of Jupiter: Io, Europa, and Ganymede. A comparison of the orbital periods in Table 1 shows that these are in the ratio of 1:2:4. The origin of the Laplace resonance and its stability is attributed to tidal interactions between Jupiter and its three resonant moons. The orbits expand in the resonance as rotational energy is transferred through tidal interaction from Jupiter to Io and passed on in part to Europa and Ganymede (see Chapters 10.08, 10.17, and 10.18). The age of the resonance and how it formed are unknown. From the amplitude of libration, it has been concluded that the resonance is relatively young. Thermal-orbital modeling, however, suggests an age of at least 2 billion years (Fischer and Spohn, 1990). The tidal interaction and, in particular, the dissipation of tidal energy are believed to be the cause of volcanic activity on Io and of a subsurface ocean on Europa (Chapters 10.08 and 10.18). This ocean is completely covered by icebergs and ice shields that may slowly move relative to each other, and the ocean may even harbor, or may have harbored, primitive forms of life (see Chapter 10.14). The mathematical theory of tidal evolution is outlined in an appendix to Chapter 10.17.

10.01.5 Composition and Interior Structure of Planets

The chemical composition of the solar system is mainly dominated by the composition of the Sun, which has 98.9% of the mass of all bodies in the system. Although the compositions of the planets are different from that of the Sun, they are related to the Sun's composition and reflect varying degrees of depletion in volatile elements (see articles in Volume 1 of the Treatise on Geochemistry, Davis (2004)). The Sun is mainly composed of hydrogen and helium but contains all the other elements found in the solar system in abundances that are characteristic of the so-called solar composition.

Hydrogen and helium are dominant in Jupiter and Saturn, which are massive enough to keep these elements in molecular form against their tendency to escape to space. The potential of a planet to keep a specific element is measured by its escape velocity, which is proportional to the square root of the ratio between the planet's mass and its radius and must be compared with the thermal speed of an element, which is inversely proportional to its molecular weight (see Chapter 10.13). Still, Jupiter and Saturn are depleted in both H and He with respect to the composition of the Sun (Chapter 10.16). In addition to H and He, Jupiter and Saturn contain substantial amounts of water (H_2O), ammonia (NH_3), and methane (CH_4), compounds collectively known as the planetary ices because of their occurrence on the surfaces of the major planets' icy satellites. In the deep interiors of the giant planets, these compounds are to be found in their supercritical forms for which there is no difference between the gaseous and liquid states. In addition to H, He, and planetary ices, Jupiter and Saturn likely have central cores of rock and iron. Models of the interior structure of planets can be constructed from a sufficiently detailed knowledge of their *figures* and their *gravity fields*. (The gravity fields of the terrestrial planets are discussed in Chapter 10.05; interior structure models of the terrestrial planets in Chapter 10.02. Chapter 10.04 discusses the rotation of these planets and how the rotation depends on interior structure. The giant planets are discussed in Chapter 10.16, while interior structures of giant planet satellites are discussed in Chapter 10.18) The construction of interior structure models is a primary task of planetary physics. Rotation deforms the planets from spheres into prolate spheroids. The flattening of both the figure and the gravity field is dependent upon the variation of density with depth. Unfortunately, these data do not allow unique models. The most accurate method of exploring interior structure is *seismology* (Chapter 10.03), the study of sound waves that travel through the interiors of planets and of global oscillations. This method is the prime method for exploring the Earth's interior to provide models of the variation of density with depth and also laterally by a technique that is known as seismic tomography (see Volume 1 of the Treatise on Geophysics). Seismometers, the instruments to record the waves, have been placed on the Moon during the Apollo program, and the interpretations of the results have provided important insights even up to the recent discovery of a solid inner core in the Moon (Weber et al., 2011). The InSight mission currently build by NASA in cooperation with CNES and DLR will place a state-of-the-art very broad band seismometer on the surface of Mars in 2016. The acoustic coupling between the surface and the atmosphere offers chances to do seismology from the orbit using cameras, infrared mappers, and ionospheric sounders. This method offers particularly good chances for Venus where environmental conditions on the surface are exceedingly demanding. The method can also be applied at the giant planets to study free oscillations and constrain interior models with these.

Models of the giant planets suggest that the outer layers of both Jupiter and Saturn are mostly H and He. H is molecular at moderate pressures but becomes metallic at pressures larger than 170 GPa (Figure 9). This pressure is equivalent to depths of about 14 000 km in Jupiter and 27 000 km in Saturn (Chapter 10.16). The phase transformation occurs at greater depth in Saturn because the pressure increases at a smaller rate in this planet due to its smaller mass. The deeper interiors of both contain the denser ices possibly in a shell above the densest iron/rock core. The cores plus the ice shells are believed to have masses that are quite uncertain, up to 45 Earth masses and radii of a few Earth radii.

Uranus and Neptune are further depleted in H and He and consist mainly of ice and the rock/iron component in addition to some H and He. Their structure, as the interpretations of the gravity field reviewed in Chapter 10.16 suggest, is not as clearly layered as those of Jupiter and Saturn. Rather, there appears to be a gradual increase of density with depth accompanied by a gradual increase of the abundance of ice at the expense of H and He followed by a gradual increase of the abundance of rock/iron at the expense of ice.

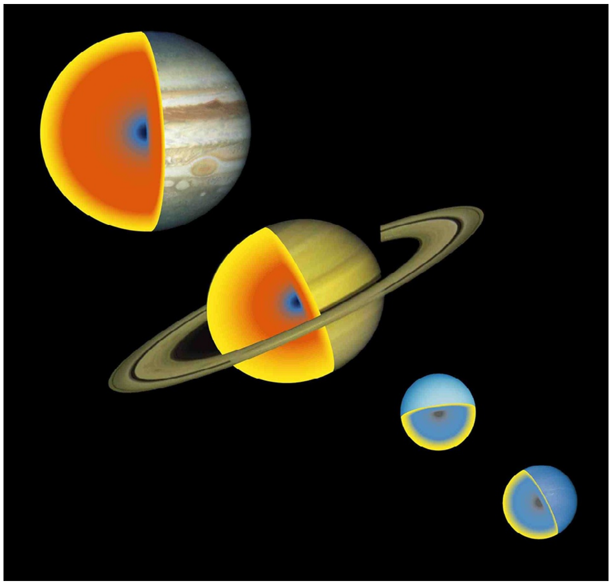

Figure 9 Interior structures of the giant planets. The yellow color indicates the molecular H and He shroud. Orange indicates the metallic H and He layers, blue is the range where the ice components are found, and the dark colors indicate the rock/iron cores. Courtesy of T. Guillot.

The major satellites of the outer solar system and Pluto also contain substantial amounts of ice, as their densities, around 2000 kg m^{-3}, suggest (see Table 1 and Chapters 10.18, 10.19, 10.20). Examples to the contrary, with substantially greater densities, are Io and Europa, the two innermost Galilean satellites of Jupiter. It is tempting to speculate that these two also started their evolution with a substantial abundance of ice. Io may have lost the ice completely and Europa to a large extent as a consequence of heating by impacts or as a consequence of tidal heating. Since Io and Europa are closer to Jupiter than Ganymede and Callisto, both the energy of impactors and the tidal dissipation rate will have been larger and may have caused the loss of water. But it is not clear that tidal heating could have provided the power needed to vaporize or partly vaporize ice layers on the two. However, it is also plausible that the temperature in the Jovian nebula from which the satellites formed was too hot at Io's and Europa's orbital distance to allow the accumulation of substantial ice shells. (The accretion of the satellites of Jupiter is discussed together with the formation of other satellites in Chapter 10.17) The Galileo mission has returned two-way Doppler ranging data that allow reasonable modeling of their interior structure (Figure 10). (See Chapter 10.18 for the icy satellites and Chapter 10.02 for Io. See also Chapter 10.23 for a brief discussion on the acquisition of radio science data.) Accordingly, Io has an iron-rich core of about half the satellite's radius and 20% of the satellite's mass with a rock shell above. Europa has a (water) ice shell about 150 km thick overlying a silicate rock shell and an iron-rich core. The nonuniqueness of interpretation of the data allows a wide range of sizes of these reservoirs, but the most likely radius values are 500–600 km for the core and 800–1000 km

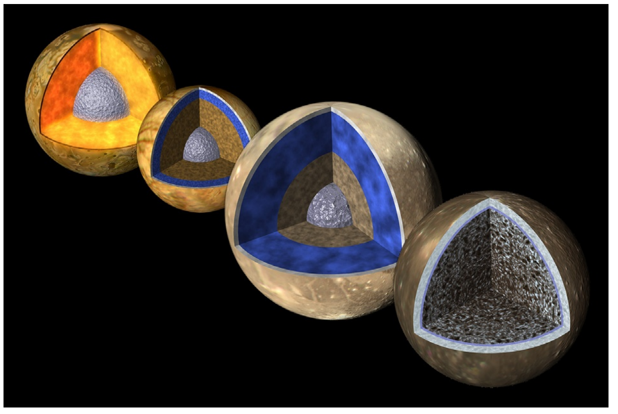

Figure 10 Interior structures of the Galilean satellites of Jupiter, Io, Europa, Ganymede, and Callisto (background to foreground).

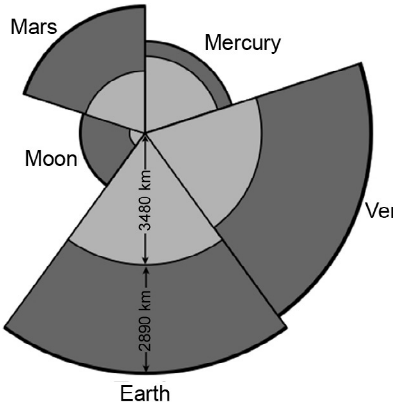

Figure 11 Basic interior structure of the terrestrial planets. To the first approximation, all terrestrial bodies possess a layered structure consisting of an innermost metallic core (light gray section), a silicate rocky mantle (dark gray layer), and a thin crust of volcanic origin, which is chemically distinct from the mantle (black layer not drawn to scale). While planetary radii are known precisely, core radii (light gray section) are not, apart from the case of the Earth's core, which consists of a liquid part surrounding a solid inner core (not drawn) with a radius of 1220 km. Reproduced from Tosi N, Breuer D, and Spohn T (2014) Evolution of planetary interiors. In: Spohn T, Breuer D, and Johnson TV (eds.) *Encyclopedia of the Solar System*. Amsterdam: Elsevier.

for the rock mantle. It is widely believed that a large part of the ice shell may actually be liquid allowing for an internal ocean. This is suggested by the results of the Galileo mission magnetic field measurements that indicate a field induced by Europa's motion in the magnetic field of Jupiter. Moreover, model calculations show that tidal dissipation in the ice may produce enough heat to keep an ocean underneath an ice lid a few tens of kilometers thick. It is also possible, depending on little known values of *rheology* parameters and thermal conductivity, that radioactive decay and tidal heating in the rock mantle may keep the ocean fluid (Chapter 10.18).

Ganymede has an ice shell about 800 km thick, a 900 km thick rock mantle shell, and a metallic core with about the same radius. The case for an iron-rich core is particularly appealing for Ganymede because the magnetometers on Galileo have detected a self-generated field on this satellite (see Chapter 10.06). The intrinsic magnetic field of Ganymede is most likely generated in an iron-rich core. The magnetic field data also suggest an induced component although the evidence for the induced field is much less strong. If the latter interpretation is correct, then Ganymede may also have an ocean of perhaps 100 km thickness at a depth of around 150 km. Since tidal heating is negligible at present, the heat that keeps the ocean molten must either be derived from radioactive decay in the rock mantle or from heat stored in the deep interior during earlier periods of strong tidal heating or even from accretion. In any case, the existence of the self-generated field is evidence enough for a molten core or, at least, a molten core shell.

Callisto is unusual among the Galilean satellites because the gravity data suggest that its interior is not completely differentiated. It is likely that there is an ice shell a few 100 km thick overlying a mixed interior of ice and rock also containing iron. It appears as if Callisto has traveled an evolutionary path different from that of Ganymede. The layering may have formed as a consequence of the slow unmixing of the ice and rock/iron components (Nagel et al., 2004). The model requires that Callisto has never been heated above the melting temperature of the ice. It is possible if not likely that the iron is no longer present in its metallic form but is oxidized to form magnetite and fayalite. The oxidation of the iron, if it occurred, would have precluded the formation of a metallic iron core. As for Europa and Ganymede, the magnetic field data suggest an induced field and an ocean for Callisto as well. Since Callisto is not in the Laplace resonance, tidal heating can be discarded as a heat source. The Cassini gravity data for the largest Saturnian satellite Titan suggest that it too is incompletely differentiated, thereby resembling Callisto, a satellite, the surface of which is so different.

The terrestrial planets (Figure 11) have ice in only modest concentrations and mostly consist of the rock/iron component. The rock/iron component in these planets, just as in Io, Europa, and Ganymede, is differentiated into a mostly iron core, a silicate rock mantle, and a crust consisting of the low-melting point components of rock (Chapter 10.02). The crust, a layer of some tens of kilometer thickness, forms by partial melting of the mantle and melt separation from the mantle. Crustal rock is still being produced on Earth by volcanic activity, but volcanic activity has mostly ceased early on Mars and Mercury (compare Chapters 10.02, 10.08, and 10.09 and Toplis et al. (2013) for discussions on Mars).

Seismology has provided us with a detailed image of the interior structure of the Earth. It is expected that future missions will provide us with similar data for the other terrestrial planets (see Chapter 10.03). The most likely candidate here is Mars and the Moon, for which mission scenarios with seismological networks have been repeatedly studied. The seismological data for the Earth show that there are phase transformations mostly at moderate pressures and depths and chemical discontinuities. The existence of these layers and their thicknesses and depths will vary among the planets. For instance, it is likely that Mercury's mantle will not have major phase-change layers simply because the pressure in this small planet does not reach the levels that it reaches within the Earth, but it may be chemically layered as is the Moon. Although the latter is even smaller than Mercury, the seismic data available suggest a layered structure.

The case of the Moon illustrates the importance of data archiving. Only recent progress in seismology has provided the tools to more fully exploit the Apollo data and has demonstrated the layering of the lunar interior including a liquid outer and a solid inner core (Weber et al., 2011). The layering of even small bodies may be due to the very early evolution of the planets. There is evidence that the planets were once covered by what is called a magma ocean, rock molten by the dissipation of the energies of infalling planetesimals during the late stages of accretion. The Moon possibly became chemically layered as the magma ocean fractionally crystallized. The Earth may have escaped that layering because vigorous convection mixed the products of fractional crystallization. The vigor of convection depends on the size of the body or layer undergoing convection.

10.01.6 Surfaces and Atmospheres

The surfaces of most planets and major satellites have been observed optically, in the visible wavelength range of the electromagnetic spectrum, by cameras onboard spacecraft. The gaseous giant planets and subgiants show the top layers of their cloudy atmospheres. Since these bodies lack solid surfaces, their surface radius has been defined to be the radius at which the atmospheric pressure is 10^5 Pa. Mercury and the major satellites except Titan show their solid surfaces. These bodies are lacking substantial atmospheres. The ability of a planet or satellite to keep an atmosphere depends on the planet's gravity (compare **Chapter 10.13**). Equating the kinetic energy of a molecule to its potential energy defines the escape velocity that is independent of the mass of the particle. Atoms, ions, or molecules move at their thermal speed – which obeys a Maxwell–Boltzmann distribution and is proportional to the square root of the ratio of the temperature to the particle mass – but can be accelerated by other processes much beyond their thermal speeds. Among these processes are impacts, photoionization, and pickup by the solar wind, hydrodynamic effects, and sputtering. The presence of a magnetic field will help to keep an atmosphere by blocking the solar wind from eroding the atmosphere. Escape due to the thermal speed (at the long end of the Maxwell–Boltzmann distribution) exceeding the escape velocity is called the Jeans escape and is most relevant for light species such as hydrogen and helium. The Jeans escape can be used for a systematic discussion in terms of planetary mass, radius, and temperature. It explains why Mercury with its comparatively small mass and high surface temperature is prone to losing an atmosphere and why massive Jupiter can bind hydrogen and helium, but it cannot explain why Titan has a substantial atmosphere, while Ganymede, similar in mass, radius, and temperature, has none. The difference may lie with Titan being rich in ammonia from which the mostly nitrogen atmosphere may have formed by photodissociation (Coustenis and Taylor, 1999 and **Chapter 10.18**). It is also possible that the difference between Titan and Ganymede lies with their early differentiation and outgassing histories. The compositions of the atmospheres of the terrestrial planets suggest that any solar-type primordial atmosphere has been replaced by a secondary atmosphere that is the result of outgassing and perhaps cometary impacts (e.g., Pepin, 2006).

The surface temperatures of the planets are mostly determined by the solar radiation and therefore decrease with increasing distance from the Sun. The actual surface temperature will depend on the reflectivity or *albedo* in the visible wavelength of the electromagnetic spectrum, on the emissivity in the infrared, and on the thermal and optical properties of the atmosphere. A useful quantity, the *effective temperature*, can be calculated assuming thermal equilibrium of the surface with the solar radiation neglecting the effects of the atmosphere. The effective temperature decreases from 445 K at Mercury to 50 K for Neptune (**Table 1**). In case of an atmosphere, the energy balance is more complicated (**Chapter 10.13**). Visible light passes through the atmosphere and reaches the surface where it is in part reflected and in part absorbed and reradiated. The infrared radiation emanating from the surface is in part absorbed by CO_2 and/or H_2O in the atmosphere and raises the atmosphere temperature. This effect, the so-called greenhouse effect, is particularly important for Venus (see also Bengtson et al. (2013)) due to the large amount of CO_2 in its atmosphere. (Other minor constituents contribute to the greenhouse effect.) Here, the surface temperature is about 740 K, about twice the value of the effective temperature. It is possible that Mars had a much denser atmosphere in its early evolution that allowed for higher temperatures through the greenhouse effect and liquid water. This would have helped to develop life on this planet (**Chapter 10.14**).

The solid surface of the Earth is partly covered by clouds in the atmosphere and partly visible depending on the extent and the pattern of the cloud cover. Venus' surface, however, is entirely covered by an optically dense atmosphere and the same is true for the Saturnian satellite Titan. Venus' atmosphere is mostly CO_2 and the atmospheric pressure is 9.2 MPa. The pressure at the bottom of Titan's atmosphere is 150 kPa. The orange-colored haze that hides Titan's surface from view is most likely smog produced by the photodissociation of methane. Venus' surface has been explored by radar observation. Titan is also explored by radar through the Cassini mission, which, at the time of this writing, is orbiting the planet and has repeatedly flown past Titan at a distance of about 1000 km. Radar and infrared mapping data have revealed a variety of morphological surface features such as lakes, dunes, and perhaps ice volcanoes suggesting a complex story of climate, geology, and atmosphere/surface interactions (e.g., Jaumann et al., 2009). On 14 January 2005, the European Huygens probe onboard Cassini successfully landed on Titan and returned the first images from its surface. The images reveal a complex surface morphology and suggest rivers and perhaps oceans (**Figure 12**).

The surface of Mars is visible most of the time. The main component of the Martian atmosphere is CO_2 but the atmospheric pressure is much smaller than Venus', of the order of 600 Pa. However, the surface may be hidden during times of global dust storms, which occur repeatedly on the timescales of a few years. Ancient river beds, outflow channels, and erosion features have been taken as evidence that the Martian atmosphere was more massive in the past and the climate was wetter and warmer (see **Chapter 10.11** for a discussion of water and ice on terrestrial planetary surfaces, in particular on Mars, and **Chapter 10.12** for a discussion of atmosphere–surface interactions). However, this view is debated and

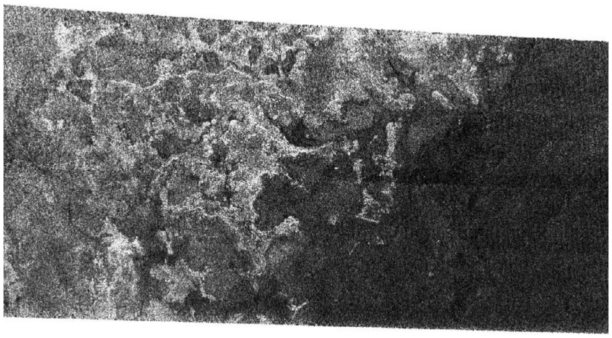

Figure 12 Synthetic aperture radar image of the surface of Titan. The bright, rough region on the left side of the image seems to be topographically high terrain that is cut by channels and bays. The boundary of the bright (rough) region and the dark (smooth) region appears to be a shoreline. The patterns in the dark area indicate that it may once have been flooded, with the liquid having at least partially receded. The image is 175 km high and 330 km wide and is located at 66 degrees south latitude, 356 degrees west longitude in the southern hemisphere of Titan.

alternative scenarios are discussed (Chapter 10.11). The findings of the Mars Exploration and Curiosity rovers clearly show that standing bodies of water on Mars were once present in which water-bearing minerals like jarosite precipitated. Water is a prerequisite for a planet to be habitable.

The pressure on Mars is close to the triple-point pressure of water (611 Pa). Thus, there is little room for liquid water except for low-lying regions such as the bottom of impact basins like Argyre and Hellas. However, melt water may exist for some time metastably on Mars and longer as its surface freezes over and it is covered by ice (see also articles in Tokano (2005) and references therein). The widely accepted early disappearance of the magnetic field of Mars (see Schubert et al. (2000) for a dissenting view) may have been partly responsible for the escape of the early Martian atmosphere (see Dehant et al. (2007) for a recent review).

The atmosphere of the Earth differs from those of its neighbors. The atmospheric pressure of 0.1 MPa is in between that of the latter two, and the main components are N_2, O_2, and H_2O.

The outer surface of Uranus is bland, greenish in color, and mostly featureless. The greenish color is attributed to methane and high-altitude photochemical smog. At the other extreme is Jupiter, whose surface features a large number of bands or stripes largely parallel to the equator interspersed with spots and vortices. Particularly remarkable is the giant red spot, a vortex that covers about 1/10 of the planetary disk (see Figure 1). These features point to a highly dynamic atmosphere. The dynamics is dominated by the rotation as witnessed by the band structure, but the vortices and spots show that these become unstable at various scale lengths. The brownish to reddish color of some of the features is attributed to ammonium hydrogen sulfide (NH_4SH), while the whitish colors are attributed to ammonium. Saturn's atmosphere is much calmer than Jupiter's, as the surface patterns suggest. These patterns are stripes similar to those in the Jovian atmosphere, but the spots and vortices are missing although wind speeds are extremely high. Neptune is similar to Uranus in that its surface is bluish in color, which is attributed to methane. It resembles Saturn in its stripe pattern. But in addition, there are a few vortices and a large spot that resemble features on Jupiter (the atmosphere dynamics of the giant planets is discussed in Chapter 10.16).

The solid surfaces of the terrestrial planets have some common features. Most prominent among these are craters that occur on a very wide range of sizes (compare Chapter 10.10). Craters are believed to be remnants of the early evolution of the planets and satellites when the young surfaces were bombarded by planetesimals during the late stage of the accretion process. This phase is often termed the phase of heavy bombardment. The distribution and density of craters on planetary surfaces are an important indication of the age of the surface or parts thereof. The older a surface is, the higher the density of impact craters. For instance, Mars shows a dichotomy in surface age (and topography) between its northern lowland and southern highland hemispheres (see Figure 8). The age difference on average is about 1 billion years. Some of the most cratered and therefore oldest surfaces in the solar system are found on Mercury, the Earth's Moon, and the Jovian satellite Callisto. There are surfaces that are saturated with craters, implying that for every new crater formed by an impact, one existing crater will be – on average – destroyed. Geologists and planetary scientists use the term 'exogenic dynamics' to characterize the processes modifying the surface by impacts. Many impact craters on terrestrial planets with atmospheres have been partly or completely destroyed by erosion. Some have been buried by extensive resurfacing in the early history of the planet. On icy satellites, long-term relaxation of the ice may cause craters to flatten out.

Planetary surfaces are also modified by erosion (Chapter 10.12) and by processes related to 'endogenic dynamics,' that is, by processes that originate from within the planet (see Chapters 10.08 and 10.09). The most prominent of the latter processes is volcanism. Volcanism has shaped at least parts of the surfaces of the terrestrial planets, and silicate volcanism is the one important element of crustal growth. Prominent volcanic features are the giant Tharsis dome on Mars, the island arcs on the Earth, and the Maria on the Moon. The surface of Venus is dotted with volcanoes and plains. As we will discuss further in the succeeding text, endogenic activity scales with the mass of a planet. This partly explains why the activity on smaller planets usually dies off earlier than on the larger planets.

A special case is the Jovian satellite Io (see Chapter 10.08 and articles in Spencer and Lopes (2006)). This satellite is similar in mass to the Moon, yet it is the most volcanically active planetary body in the solar system. Volcanic features cover its surface but impact craters have not been detected. This shows that the surface is very young and is permanently renewed. The surface is to a large extent covered by allotropes of sulfur and by sulfur dioxide. These deposits cause the yellow to whitish color of the surface. The dark spots are likely volcanic vents and sulfur lakes. Sulfur deposits are often found associated with volcanic structures on the Earth. The high surface temperatures of up to 1600 K indeed suggest silicate volcanism. The reason for the unusual activity on Io lies with an unusual heat source: Io is flexed by tides raised by its massive primary Jupiter. The deformation energy that is dissipated as heat is sufficient to make it more volcanically active than the largest terrestrial planet, the Earth. A further measure of the enormous energy that is dissipated in Io's interior is the surface heat flow of 2–3 W m^{-2}, 20–30 times larger than the surface heat flow of the Earth. Io's heat flow is rivaled only by Jupiter, which radiates about 5.4 Wm^{-2}. The heat flow through the southern polar cap of Enceladus is also exceptionally high, however (Figure 13).

The Earth features a style of endogenic activity, plate tectonics, that, according to our present knowledge, is unique to the Earth (chapters in Volumes 6 and 7 and Chapters 10.08 and 10.09). Plate tectonics involves the continuous production of basaltic crust along volcanically active linear ridges at the bottom of the oceans. Prominent examples are the Mid-Atlantic Ridge and the East Pacific Rise. It also involves subduction of this crust underneath island arcs and continental margins. Prominent examples are the islands of Japan for the former and the western continental margin of South America for the latter. These subduction zones are the loci of most of the seismic activity of the planet. Both processes cause the Earth's surface to be divided into seven major plates that drift across the surface. Plate tectonics is driven by extremely slowly circulating convection currents in its deep interior. Although convection is not unique to the Earth but is expected to occur in the silicate rock shells of the other terrestrial planets as well, this feature is usually hidden underneath a thick stagnant lid (see Section 10.01.7 and Chapters 10.08 and 10.09). The reason why convection in the Earth extends to the surface and includes the crust at the bottom of the oceans is not entirely known. It is speculated that this is caused by the presence of water and there may even be links between plate tectonics and life (e.g., Höning et al., 2014; Sleep et al., 2012). In any case, plate tectonics causes the surface of the ocean basins to be completely renewed on a timescale of a few hundred million years. Plate tectonics together with erosion even incorporates recycling of the more stable continents.

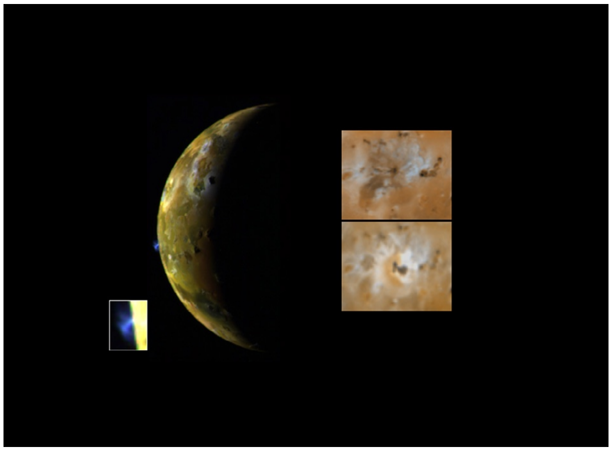

Figure 13 Volcanic plume on Io imaged by the Galileo spacecraft. The blue-colored plume extends to about 100 km above the surface. The blue color is consistent with the presence of sulfur dioxide gas and sulfur dioxide snow condensing as the volcanic gas in the plume expands and cools. The images on the right show a comparison of changes seen near the volcano Ra Patera since Voyager in 1979 (top) and Galileo in 2001 (bottom). An area of 40 000 km^2 was newly covered with volcanic material between the two observations. NASA/JPL.

Material denudated by erosion from the continents is transported to the ocean basins where it is incorporated into the plate tectonics cycle. That loss of continental material is balanced by the production of continental rock through volcanic activity. The cycling of material between the interior and the atmosphere is an important element of the carbon silicate and other elemental cycles that help stabilize the climate of the planet (cf. Chapter 10.13). It has been speculated that Mars and Venus went through phases of plate tectonics very early in their histories (see Chapters 10.08 and 10.09). However, while this is a possibility, it remains a speculation. Although plate tectonics seems to be unique to the Earth, there are other processes of crust recycling. These involve subcrustal erosion as is assumed for, for example, Io and the foundering of the surface as has been speculated to have happened on Venus at the sites of some coronae Ghail (2002) even proposed a mechanism of "buouyant plate tectonics" for present day Venus.

The best studied example of a single-plate planet is Mars. Three instruments on board spacecraft have recently allowed major progress for detailed studies of the tectonics of this planet (Chapter 10.09). The first is the laser altimeter MOLA on Mars Global Surveyor, the second is the HRSC camera on Mars Express, and the third is the HiRISE camera on the Mars Reconnaissance Orbiter. While MOLA has stopped operating as an altimeter in 2001 after covering the entire surface due to the finite lifetime of its laser pump diodes, HRSC has covered almost the entire planet at a resolution of 10 m px^{-1} at the time of this writing and HiRISE has supplemented this data set with images down to 30 cm px^{-1}. One major achievement by these instruments is the establishment of an accurate geodetic network of surface features.

The topography of Mars exhibits a clear dichotomy that divides the surface into a southern highland hemisphere as heavily cratered as the lunar highlands and rising several thousands of meters above the zero level and a northern lowland hemisphere that lies well below the datum. The origin of the dichotomy is ascribed variously either to long-wavelength mantle convection, to impacts, to postaccretional core formation sweeping up most of the crustal material into one large protocontinental mass, or to combinations thereof (see discussion in Chapters 10.08 and 10.09). The highlands cover about 60% of the planet, including almost all of the southern hemisphere. The surface has survived since the heavy bombardment prior to 3.8 billion years ago with only minor modifications and thus records the early history of the planet.

Venus whose surface has been mapped by the Magellan and Venus Express missions with radar and in the infrared is another well-studied single-plate planet. Its hypsometric curve, however, has a single peak rather than having double peaks as those of Mars' and the Earth's curve. Venus has no distinct volcanic center such as Mars but a prominent rift system, the Beta, Atla, and Themis Regiones. Volcanic features include a large number of comparatively small shield volcanoes that are apparently randomly distributed, volcanic plains, and about 500 coronae – circular large-scale volcanotectonic features. It is widely accepted that most coronae form above upwelling plumes (Smrekar and Stofan, 1999). The comparative paucity of impact craters on the surface has been used to suggest that Venus was resurfaced about 500–700 million years ago by a volcanic event of global scale (Schaber et al., 1992;

McKinnon et al., 1997). This event was followed – as the model suggests – by almost no volcanic activity. The early conclusions from the cratering record have been challenged by, for example, Hauck et al. (1998) and Campbell (1999). According to these authors, the cratering record allows a variety of interpretations in terms of volcanic resurfacing including a global decrease in time in the rate of volcanic activity.

10.01.7 Energy Balance and Evolution

Although the evolution of the planets seems to have followed some common general lines, there are significant differences among individual planets. The evolution of the terrestrial planets and moons is discussed in **Chapter 10.08** and of the icy satellites in **Chapter 10.18**. The evolution of the giant planets is discussed in **Chapter 10.16**. The early evolution of all planets has been dominated by impacts from the debris left over from planet formation (**Chapter 10.10**). Moreover, isotopic evidence suggests that the planets differentiated by iron-core formation early within a few tens of million years (e.g., Kleine et al., 2002). Of course, the evidence for Mars rests with the very well-founded assumption that the so-called SNC meteorite*s* are in fact rocks from Mars. It is further believed that the planets started hot, heated by the energy deposited in the interior during accretion and by heat dissipated upon differentiation – the terrestrial planets heated to near-melting temperatures of their rock components and to temperatures well above the iron-melting temperatures. The planets then cooled from this initial hot state and the cooling drove their evolution. It is possible, for instance, to explain the present rate of infrared radiation from Jupiter simply by cooling. This is not possible for Saturn, for which an additional heat source is required. The continuing gravitational settling of helium may provide enough energy to explain the observed present luminosity of this planet. The energy balances of most other planets and satellites seem to invoke both cooling and heat generated by the decay of radioactive elements, in particular of uranium, thorium, and potassium (the isotope ^{40}K), which generates 4–5 pW kg^{-1} of rock. A further possibility is tidal heating, but this mechanism seems to be relevant at present only for Io and, perhaps, Europa and Enceladus.

It should be stressed that the energy balances of the planets are not very well known. A crucial quantity, the surface heat flow – or the intrinsic luminosity – is known for the giant planets, which radiate enough energy in comparison with the solar insolation incident on their surface that their luminosity can be measured from orbit. This is also true for the heat flow from the volcanic hot spots on Io, but there is a major uncertainty concerning the flow through the remainder of the surface. Nevertheless, the enormous luminosity of this satellite of at least 1.5 W m^{-2} is remarkable. Another remarkable value is the roughly 6 GW (Spencer et al., 2006) radiated from the southern hemisphere of Enceladus. On the terrestrial planets and the other satellites, as far as we know them today, the surface heat flow must be measured *in situ* (compare **Chapter 6.05** by Jaupart). This measurement requires a borehole deep enough to avoid the influence of the daily and seasonal temperature variations of the atmosphere. This difficult measurement has been done at many locations on the Earth and at two sites on the Moon. The InSight mission scheduled for launch to Mars in 2016 will attempt a surface heat flow measurement. In **Table 1**, we list the specific luminosities or surface heat flows for the planets and satellites for which the heat flows and luminosities have been measured. Since heat production depends on mass, we have divided the heat flows by the masses. A comparison of the entries in **Table 1** shows that the Earth and the Moon are within a factor of two of the specific radioactive heat production rate of rock suggesting that about half of the heat flow can be attributed to cooling and half to heat production. This conclusion is known to hold for the Earth even on the basis of more specific data and more careful energy balances (compare **Chapters 9.08** and **9.09**). An educated prediction of the heat flows for Mars, the other terrestrial planets, and the major satellites could be based on the aforementioned observation. The specific luminosity of Uranus also is close to the radiogenic heat production rate per unit mass of rock given in the preceding text. It is likely that Uranus has a substantial rock core in which that heat is generated.

Neptune, Io, and Europa stand out in comparison with the terrestrial planets. In the case of Neptune, the extra heat flow has been attributed to cooling and whole-planetary contraction. For Io, the extraordinary large heat flow is with little doubt due to tidal heating although ohmic dissipation of energy carried by electric currents between Io and Jupiter is sometimes quoted. The value for Europa in **Table 1** is highly uncertain and has been derived from indirect observation of tectonic surface features (e.g., Ruiz, 2005). If the estimate were correct, then tidal heating would be the best explanation for its value. However, equilibrium models of tidal heating arrive at values smaller by about one order of magnitude (cf. **Chapter 10.18**).

The dynamic or geologic, chemical, and magnetic evolutions of the planets are mainly governed by their thermal evolution. Planets can be regarded as heat engines that convert heat into mechanical work and magnetic field energy. Mechanical work will be performed by the building of volcanic structures and mountain belts, for example, or during the movement of plates over the Earth's surface (cf. Volumes 6 and 7). The agent for these processes is convective heat transport, a mode of heat transport that involves movement of hot material to cold surfaces and cold material to hot surfaces (see, e.g., Schubert et al. (2001) and **Chapter 7.02**). Convection is driven by sufficiently great temperature differences across a layer or planetary shell and occurs if the layer is unstable against convective overturn. Convection is possible even in the solid rock mantles of the planets because of the long timescales that are involved. On geologic timescales of tens to hundreds of millions of years, rock behaves as a very viscous fluid. The convection flows at speeds of centimeters per year still transfer more heat than conduction. This is in part due to the enormous masses involved in the flow and in part due to the very low thermal conductivity of rock. There is even a thermostat principle at work in the silicate and ice shells (the Tozer principle) (Tozer (1967)) since the viscosities of these solids are strongly temperature-dependent: If the flow is too slow to transport the heat, the temperature will increase until the flow is strong enough to transfer the heat. If the temperature is so high such that the convective heat transfer rate largely outbalances the heat production rate or the heat flow into the layer from below, then the temperature will decrease, viscosity will increase, and convective vigor will decrease. Convection is extensively discussed in Volume 7. The thermal evolution of the Earth is discussed in Volume 9. Chemical differentiation is mostly a consequence of melting and the density difference between melt and solid that will again lead to movement. A particularly good example for the chemical differentiation of the terrestrial planets that occurs in addition to core formation is the growth of the crust through volcanic activity (see Chapter 10.08). However, secondary crustal growth likely involves much longer timescales than core formation.

The material in the crust is derived from partial melting of the mantle. Partial melting is possible when a material of complex chemistry lacks a simple melting temperature. Rather, in materials such as rock (and ice of complex chemistry), there will be a temperature termed the solidus temperature at which the material begins to melt. The melt will consist of that component of the entire assemblage that has the lowest individual melting point. The remaining solid will be depleted in that component. As the temperature rises, other components begin to melt until, finally, at the liquidus temperature, the whole assemblage will be molten. In the rock component of planets, the component to most easily melt is basalt. Once a basalt liquid forms in the interior of a planet, it will tend to rise to the surface by virtue of its lower density than the density of the remaining rock. In a water–ammonium ice, the low-melting component will be a water–ammonium mixture of a particular composition. In most cases, the melt is likely to be produced within the top few 100 km of a planet by a process termed pressure-release melting. Pressure-release melting occurs when the melting point gradient is steeper than the average temperature gradient. Relatively hot uprising convection currents will emanate from depths where the temperature is below the melting temperature. Upon rising to the surface adiabatically (i.e., with no or little heat exchange), the flow will hit a depth where the solidus temperature is exceeded and partial melting will ensue. Because melt is usually much more compressible than solid, there will be a pressure and a depth below which melt will no longer be buoyant to rise to the surface but may actually sink towards the deeper interior. Melting in the interior of planets requires a heat source. This source is mainly the heat that is stored in the planets during accretion and the heat that is generated by the decay of radiogenic elements. Both are finite reservoirs that will not be replenished. That is why endogenic activity decreases with time.

Since melting occurs when hot material from the deep interior rises in convection currents, the surface distribution of volcanism is speculated to give an indication of the planform of the convection underneath. The timing of the volcanic activity can be used as a guide for assessing the time evolution of the convection. For instance, the differing surfaces of Mars and Venus may have recorded differing planforms of mantle convection and flow history. The dominance of Tharsis on Mars (compare Figure 8) suggests that there is or that there once was a giant superplume, a very large upwelling underneath this volcanic dome. The geologic evidence suggests that Tharsis formed early but that volcanic activity has been ongoing to the recent past (cf. Chapter 10.08). This long-term stability of Tharsis as the major center of volcanic activity on Mars is puzzling, the more so since model calculations suggest that the superplume's lifetime should be not much more than one billion years. Although Tharsis apparently formed early, the photogeologic evidence also suggests that volcanic activity on Mars started globally and retreated to Tharsis over time and may then have decreased in vigor. On Venus, there is an indication that the volcanic activity has been global even recently (on geologic timescales, recently involves the past 10–100 million years). There is further (albeit debated) evidence for a global volcanic resurfacing event a few hundred million years ago. This suggests that the thermal history of a planet may have involved episodes of more and of less activity that most likely were linked to episodes of greater and less vigor of convection. Volcanism may also have influences on a hydrosphere and atmosphere by degassing the deep interior of volatile elements. At the time of strong volcanism, an increase of atmospheric gases or water on the planet's surface is expected.

There is a fundamental difference between the evolutions of the Earth and most other terrestrial planets and major satellites that is related to the occurrence of plate tectonics on the Earth (discussed in Chapters 10.08 and 10.09). Since the viscosity or, more generally, the rheology of rock is strongly temperature-dependent and since the surface temperatures of these planets are much lower than the melting temperatures, it follows that there must be outer layers that are comparatively stiff. These layers are termed the lithospheres. Usually, the lithospheres are connected lids that are pierced here and there by volcanic vents. Convection currents flow underneath the lithosphere and deliver heat to the base of the lid through which it is then transferred to the surface by heat conduction. Since these lids are stagnant, this form of tectonics is often termed stagnant-lid tectonics. On Earth, however, the lithosphere is broken into seven major plates that move relative to each other driven by the convective flow underneath. Volcanic activity occurs along some plate margins and results in the growth of plates. Other plate margins are destructive and are loci where plates are forced under their own weight to subduct into the mantle. There is no convincing evidence that plate tectonics occurs on any other planet or satellite although it has been speculated that it may have occurred on early Mars and on the Jovian satellites Europa and Ganymede. There is evidence in the images of the surface units of lateral movement on these two satellites. The Jovian satellite Io does not seem to be undergoing plate tectonics. However, its present resurfacing rate, which is larger than the resurfacing rate of Earth, calls for some recycling of Io's crust volcanic material with the underlying mantle. This recycling probably occurs through delamination of the base of the crust. Delamination of the base of the crust may have been or may be operative on other planets such as Venus (Figures 14 and 15).

The interior evolution of the gaseous planets is even less well constrained. It must be assumed that the dynamics of their atmospheres is related to the vigor of convection underneath, but the solar insolation also matters. Thus, models of the evolution are mostly constrained by their luminosities, but as we have seen in the preceding text, these leave enough puzzles

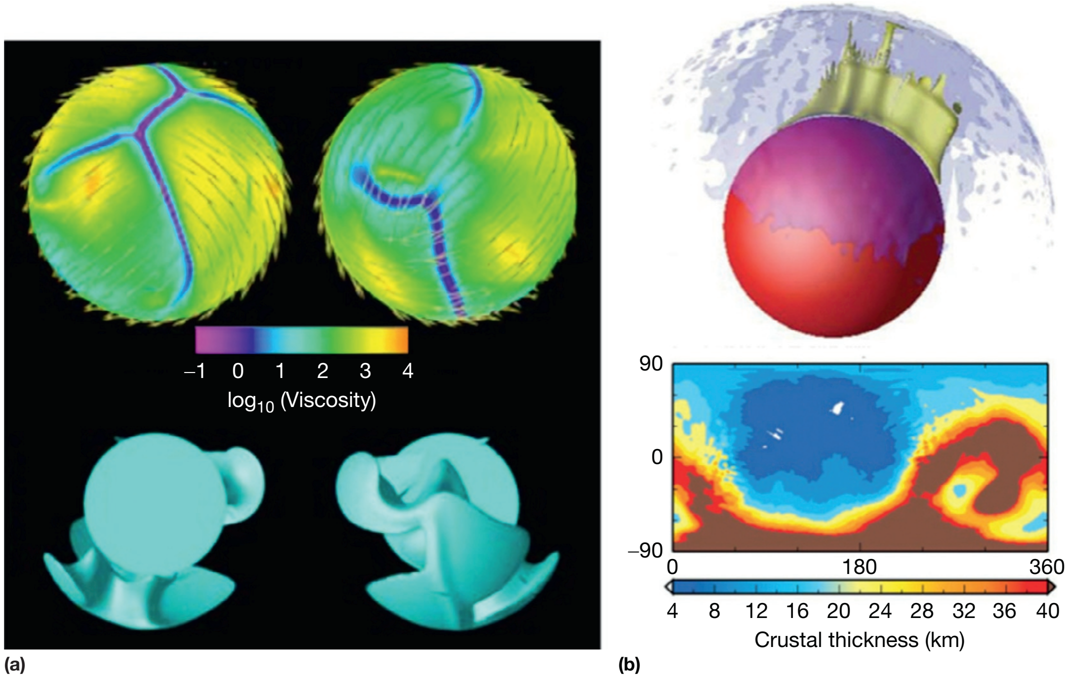

Figure 14 (a) Numerical simulations of mantle convection with plate tectonics in a three-dimensional spherical shell. The top line shows two views of the viscosity field at the surface. Rigid plates moving with uniform velocity are separated by both divergent and convergent narrow boundaries with low viscosity at which plates form and are recycled into the mantle. The bottom line shows temperature isosurfaces representing the subduction of cold downwellings (after van Heck H and Tackley PJ (2008) Planforms of self-consistently generated plate tectonics in 3-D spherical geometry. *Geophysical Research Letters* 35: L19312, http://dx.doi.org/10.1029/2008GL035190). (b) Numerical simulation of the formation of Mars' crustal dichotomy by mantle convection. The top panel shows a hot upwelling plume (yellow) rising from the core–mantle boundary (red) able to generate a large region of partial melt (light blue). Migration of the melt towards the surface leads to the formation of a crust (**Section 10.01.7**) whose thickness in the southern hemisphere is much greater than in the northern one (bottom panel) (after Šramek O and Zhong S (2012) Martian crustal dichotomy and Tharsis formation by partial melting coupled to early plume migration. *Journal of Geophysical Research* 117(E01005), http://dx.doi.org/10.1029/2011JE003867).

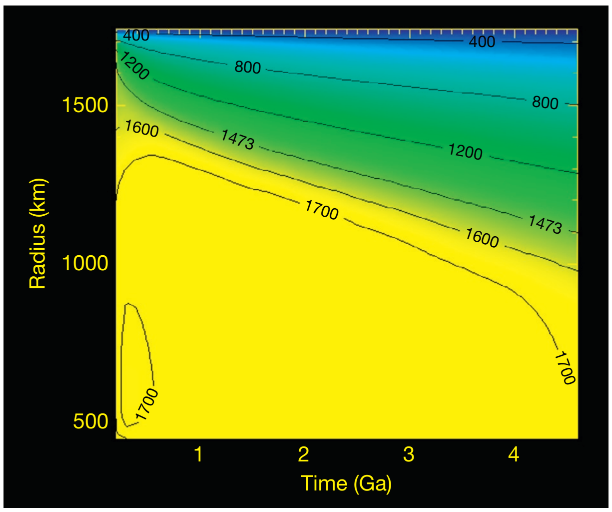

Figure 15 Growth of a stagnant lid in the lunar mantle. The growth of a cold, thermally conductive lid on top of the lunar mantle is shown as resulted from convection model calculations by Konrad and Spohn (1997).

to be solved as witnessed by the differences between the luminosities of Neptune and Uranus.

10.01.8 Magnetic Fields and Field Generation

Mercury, Earth, and the giant planets have largely dipolar magnetic fields that are produced by dynamo action in their interiors. Of the major satellites, only Ganymede is known to produce a magnetic field. The Galileo and Cassini data suggest that neither the other Galilean satellites of Jupiter – in particular not dynamic Io – nor the big Saturnian satellite Titan have self-generated fields. The magnetic fields of these planets and moons are described and discussed in Chapter 10.06. The physics of planetary dynamos is discussed in Chapter 10.07. Chapter 10.08 discusses the generation of the magnetic fields of terrestrial planets in the context of their thermal evolution, and Chapter 10.16 the generation of magnetic fields in the giant planets. Chapter 10.18, finally, discusses the magnetic fields of outer solar system satellites. Planetary magnetism is also extensively reviewed and discussed in Christensen et al. (2011).

The icy Jovian satellites Europa, Ganymede, and Callisto are surrounded by magnetic fields that vary along with their movement through the magnetosphere of Jupiter. These fields are, therefore, interpreted to be induced in electrically conducting layers in the satellites' interiors as the satellites orbit Jupiter and are subject to a time-varying magnetic flux. The geometry and the strengths of these fields suggest that the conducting layers are at depths of some tens to a few hundred kilometers and are most likely salty oceans. No evidence for an induced field at Titan has been reported. Thermal considerations suggest that these oceans are feasible (Chapter 10.18). Ganymede is agreed to have an induced field on top of its permanent magnetic field.

Self-sustained magnetic fields are generally thought to be enigmatic to planets, part of their interior evolution during which thermal (and potential energy) is converted into mechanical work and magnetic field energy. The Mars Global Surveyor data have confirmed this hypothesis by showing that Mars has a remanently magnetized crust. This crust must have been magnetized during an epoch when Mars was generating its own magnetic field. Interestingly, it is mostly the oldest crust units that are magnetized. It has long been speculated that Mars, Venus, and the Earth's Moon once produced magnetic fields by dynamo action in their cores (e.g., Stevenson et al., 1983). The observation of magnetized rock on the surface of Mars now confirms this general notion (Figure 16).

The dynamo mechanism (e.g., Chapter 10.07 but see also Chapters 10.08 and 10.16 and chapters in Volume 8, in particular Chapter 8.03) invoked to explain magnetic field generation is similar among the planets although the source regions differ between the terrestrial planets and major satellites, the giant planets, and the subgiants or ice giants, Uranus and Neptune. Required is an electrically conducting and fluid region that undergoes turbulent flow driven by thermal or compositional buoyancy. If a magnetic field exists, for instance, the external magnetic field of the Sun or a primary planet, a field will be induced in the source region. The flow will distort the field lines and thereby generate magnetic field energy by induction. If the flow satisfies certain conditions, the interference between the generated field and the preexisting field can be constructive, powering and maintaining a magnetic field against dissipative losses.

For the terrestrial planets and the major satellites, the candidate source regions are the fluid metallic cores or outer core shells. Because of their low viscosities (around 1 Pa), these regions are often unstable with respect to thermal convection. The stability is generally accepted to depend on the rate of heat removal from the core by the mantle convection flow. If the latter is too low (below about 10 mW m^{-2} but depending on pressure), the heat flow in the mantle can be balanced by heat conduction in the core, and the core will be stably stratified. Dynamo action is then not possible. If the heat flow into the mantle is larger than the critical value, convection in the core must be invoked to balance the heat removal rate. The situation becomes more favorable for convection in the core if the core begins to freeze and to grow a solid inner core. First, latent heat liberated upon inner core growth will help to power the dynamo. Second, the core may contain light alloying elements such as sulfur and/or oxygen. The light alloying elements expelled from the solid inner core can drive (chemical) convection and a dynamo very effectively because this dynamo will not be subject to a Carnot efficiency factor that by the second law of thermodynamics limits the efficiency at which a heat engine can do work. In this model of the chemical dynamo, the melting curve is always steeper than the core (adiabatic) temperature profile. It is believed that the cooling of the core and growth of the inner core drive the Earth's core dynamo.

A new class of chemical dynamos described in Chapter 10.08 has been discussed in recent years in particular for terrestrial planets and satellites smaller than the Earth. Examples are the Earth's Moon, Ganymede, Mercury, but possibly even Mars.

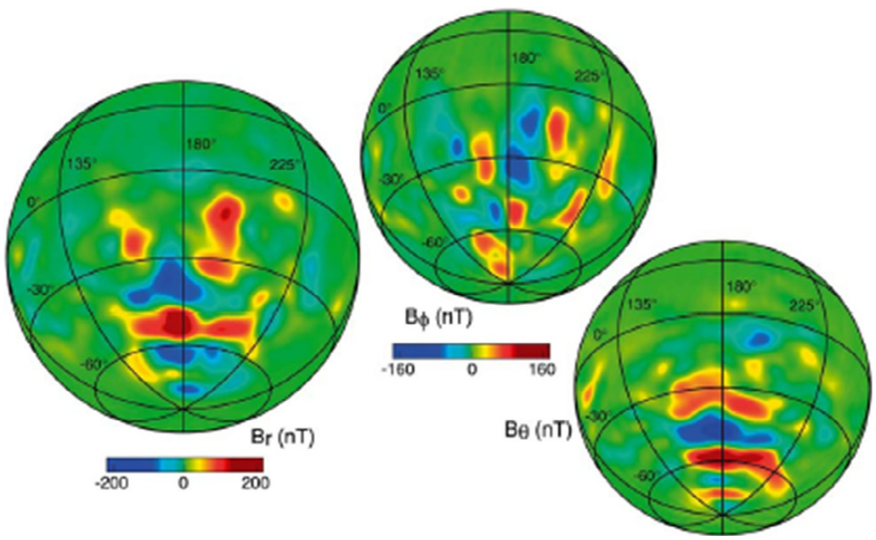

Figure 16 Orthographic projections of the three components of the magnetic field in spherical coordinates (r, radius; Θ, latitude; and Φ, longitude) onto the surface of Mars at a nominal mapping altitude of 400 km. Reproduced from Connerney JEP, Acuna MH, Ness NF, Spohn T, and Schubert G (2004) Mars crustal magnetism. *Space Science Reviews* 111: 1–32.

The new model is based on the observation that the iron-rich core alloy melting relations may be more complicated than in the older models where the application of the Lindeman melting law warranted that the melting was steeper than the adiabat. In particular, it is possible that the melting curve is less steep than the adiabat and the gradient may even be negative. In both cases, the core will not freeze from the center outward, but iron snow will form in the outermost parts of the core. As the iron snow sinks into the core, it will form a partially molten layer in which solid iron and melt are in equilibrium underlain by a layer of melt enriched in iron with respect to the average composition of the core alloy. The latter layer will be gravitationally instable and drive chemical convection and possibly a dynamo. As the core cools, the partial melt layer (with neutral buoyancy) will grow and eventually the dynamo becomes frustrated. The model may be able to explain why the Moon can have a solid inner core at present without generating a magnetic field.

Dynamo action in a terrestrial planet thus depends on the efficiency of mantle convection, on the composition of the core, and on the core material phase diagram. Plate tectonics is very effective at cooling the deep interior of a planet because the cold plates sink deep into the mantle. It is therefore conceivable that the Earth's core has been cooling to temperatures below the liquidus of its core alloy. Venus, a planet of similar size, appears to have been lacking that efficient cooling mechanism and appears to be cooling by convection underneath a stagnant lid. Consequently, the heat flow from the core has become subcritical over time, and the core may have not been cooling enough to reach liquidus temperatures. As a consequence of this and the general decline of mantle convective vigor, the core became stably stratified and a possible early dynamo ceased to operate. A similar scenario is possible for Mars, but the iron snow model may provide an alternative explanation for an early dynamo that ceased to operate. It would be interesting to see whether or not Mars has a solid inner core. There is little hope to find remanently magnetized crust on Venus should landing missions be able to overcome the forbidding operating conditions on this planet. The surface temperature of around 450 °C is above the temperatures at which candidate minerals become remanently magnetized. The Moon has a remanently magnetized crust and may be another good candidate for the iron snow model. Alternative explanations for the recorded remanently magnetized crust units invoke plasma clouds generated by major impacts. Some of the strongest magnetic anomalies (albeit weak in comparison to Mars) are suspiciously located at the antipodes of major impact basins. Mercury differs from both the Moon and Mars because of its extraordinarily large core and thin rocky mantle. The thin mantle should be quite effective, as model calculations suggest, at removing core heat and driving a dynamo that produces a magnetic field, albeit weak. The weakness of the field has been discussed to be difficult to explain with conventional dynamo theory (compare Chapter 10.07).

Dynamo action in the gaseous giant planets (Chapters 10.06, 10.07, and 10.16) most likely happens in the regions where hydrogen becomes metallic due to the extraordinary large pressure. The magnetic fields of these planets are most likely powered by the energy that has been stored in the planets during accretion and that is removed by convection. The transition to metallic hydrogen occurs at a depth of roughly 15 000 km in Jupiter and almost twice as deep in Saturn. The larger depth of dynamo action may partly explain why Saturn's field is much more ideal in terms of a dipole than Jupiter's. In the source regions, the fields are likely to be very complex and not at all similar to a dipolar planetary field. The field in the source region can be thought of as a superposition of many multipolar fields of various amplitudes. Distance from the source region affects the higher-order multipolar fields much more strongly than the lower-order fields. If n is the order ($n=2$ for a dipole), then the strength of the field components decreases with radial distance r from the source regions as r^{-n}. Thus, a planet with a comparatively deep source region should feature a more ideal dipolar field as compared with one where the source region lies less deep. This does not explain entirely the characteristics of the Saturnian magnetic field for which screening by helium rain has been additionally invoked.

The pressure in the interiors of the subgiants Uranus and Neptune is not sufficient to cause a transition to metallic hydrogen. Rather, it is believed that there are ionic oceans at relatively shallow depths in these planets. In these ionic oceans, the magnetic fields can be generated. The fields of the subgiants have been measured only by a single flyby each and are therefore not well known. Nevertheless, it appears that these fields are very complex in their topologies. This fits in nicely with the idea of the fields being generated at relatively shallow depths.

10.01.9 Origin of the Solar System

The origin of the solar system is not covered in this book. The reader is referred to Volume 1 of the Treatise on Geochemistry (Holland and Turekian, 2003). A most recent overview can be found in Chambers and Halliday (2014), and more detailed recent reviews in the Protostars and Planets V (Reipurth et al., 2007) and VI collection of articles (Beuther et al., 2014). The formation of satellites is discussed in Chapter 10.17.

The basic elements of the theory of the origin of the solar system are similar to the Kant–Laplace hypothesis of formation from a gaseous nebula, but the details are debated. It is widely accepted that the nebula collapsed to form a central mass concentration – the protosun – surrounded by a spreading thin gaseous disk. The most widely held view postulates that temperature in the inner part of the nebula soon became low enough – about 1500 K – to allow the condensation of silicate and iron grains. In the outer solar system, beyond about 5 AU, temperature became so small as to allow condensation of ice phases. Within the first few million years, these solid grains agglomerated to form bigger grains that, in turn, agglomerated to form even bigger ones. The cascading scenario of ever bigger and fewer planetesimals led to the formation of solid bodies the size and mass of the terrestrial planets in the inner solar system and a few tens of Earth masses in the outer solar system. In the outer solar system, these solid protoplanets became the cores of the giant planets. Growth of the solid protoplanets slowed to almost a standstill as the solid matter became exhausted, at least in the feeding zone of the protoplanets that is determined by the competition between a protoplanet's gravity and the gravitational pull of the Sun. At about one Earth mass, a protoplanet begins to accrete gas onto its surface. The heat dissipated during the influx of planetesimals was radiated from the surface of the forming gaseous envelope. After the influx of solid matter ceased, the protogiant planets began to cool and to contract, thereby increasing its potential to accrete more gas. The growing mass led to even more accretion and runaway growth was set in place. Runaway accretion came to a halt when a gap formed in the nebula around the planet. This gap can form as a consequence of the competing gravities of the protoplanet and the Sun and as a consequence of the finite viscosity of the gas.

A relatively new feature of solar system formation models is that giant planets can move inward (and outward) through interaction with the nebula, thereby perturbing the orbits of smaller planetesimals and protoplanets. The popular Nice model (named after the city where its basic elements were developed; Gomes et al., 2005; Morbidelli et al., 2005; Tsiganis et al., 2005) includes the effects of resonances between the orbits of the giant planets and offers explanations for the relative small mass of Mars along with late heavy bombardment and the formation of the asteroid belt.

The general model offers an explanation of the grand features of the solar system: The inner planets are solid and of refractory composition because the temperature in the inner nebula favored the condensation of refractory phases. The inner planets are small because the feeding zone was smaller with less mass of solid particles. The bigger of the inner planets – the Earth and Venus – have atmospheres with masses as expected although the present atmospheres are not likely to be the primordial ones, which were lost and replaced by degassing of the interiors. In the outer solar system, we find planets with cores of some ten Earth masses and massive gaseous envelopes. In addition, we find wealths of satellite systems witnessing accretion in orbit around the growing protogiants. There are a number of features that can be explained by fine-tuning the theory, for instance, the differences between Jupiter and Saturn and Uranus and Neptune but this will not change the grand picture.

Giant impacts, a mechanism of great importance for some bodies, are the final events of the accretion scenario. Giant impacts are thought to have caused the formation of the Moon (e.g., Canup, 2004; Galimov and Krivtsov, 2012) and may be responsible for the high density of Mercury. A giant impact is a collision of almost grown protoplanets. For instance, it is thought that the Moon formed after a Mars-sized protoplanet hit the proto-Earth. The outer layers of the Earth vaporized during the impact, and the Moon formed from the condensed vapor cloud. Among the arguments for the giant impact hypothesis for the formation of the Moon is the geochemical closeness of the bulk Moon to the Earth's mantle. Mercury may have suffered a similar giant impact that removed a substantial part of the original planetary mantle (e.g., Benz et al., 2008). This may explain why present Mercury has a comparatively big core and a thin mantle and thus an extraordinary large average density. A Moon did not form around Mercury because of the closeness to the Sun. Giant impacts may also affect the rotation, and it has been proposed that Venus' retrograde rotation may thus be explained. Similar explanations have been brought forward for Uranus.

10.01.10 Concluding Remarks

With this overview, I have tried to take the reader on a tour through the solar system and provide the stage for the detailed articles in the remainder of the volume. The planets are largely of solar composition but differ in their depletion in volatile elements. The degree of depletion increases with decreasing mass and with decreasing distance from the Sun with Jupiter being closest in composition to the Sun and with the terrestrial planets being mostly depleted in volatile elements. The planets are mostly internally differentiated with the heavy elements tending to be found near the center and the most volatile elements near the surface. The gaseous planets present the observer with their top layers of their envelopes, while most of the smaller planets present their solid surfaces. Relicts of very early impacts characterize these surfaces to varying degrees. But these surfaces also reveal the signs of volcanic and tectonic activity and erosion. In general, the smaller the solid-surface planet, the more characterized is its surface by early impacts. The volcanic activity and the magnetic fields of those planets that have self-generated fields are due to convection currents in their interiors driven by cooling and by heat generated by radioactive decay or by the dissipation of tidal energy. The source regions of the magnetic fields are metallic or, in the cases of Uranus and Neptune, ionic fluids. Life has had a chance to develop on Earth but possibly also on Mars and Europa. Finding evidence for life outside Earth is a challenge for the future. The bio–geo system of Earth may be a self-regulating system that sets this planet apart from its sister planets. Its position in the solar system, together with a moderate greenhouse effect in its atmosphere, has provided the planet with habitable temperatures and with an atmospheric pressure allowing for liquid water. Life plays an important role here by removing CO_2 from the atmosphere and moderating the greenhouse effect. Water, on the other hand, may be instrumental for plate tectonics since its effect on the rheology may pave the way for surface plates to subduct to the deep interior, thereby cooling the core. The cooling of the core is speculated to be instrumental for magnetic field generation. The magnetic field is a feature that the Earth's brother and sister planets Mars and Venus lack. The magnetic field protects the environment, and thereby life, from radiation. It will be important to see whether or not life developed in the less friendly environments of Mars or even Europa or early Venus.

The future in scientific planetary exploration on the one hand certainly lies with systematic remote sensing with long-term orbiters. These missions will create data volumes that require either long-term operation and communication or significantly increased data acquisition and transmission rates. On the other hand, *in situ* exploration with rovers and planes and networks of surface stations with long operational lifetimes to observe the atmospheres and the interiors will be important. A further challenge for the future will be the return of samples from planets and small bodies. The sophistication of terrestrial laboratories cannot simply be transferred to other planetary surfaces given the scarcity of resources such as mass and power on spacecraft. The technological challenge remains miniaturization. Although many planetary worlds are forbidding to human exploration and require the operation of robots (think of Venus), it is probably true that the human explorer cannot be topped in its ability for *in situ* scientific study and operation of hardware.

Closer to home, the challenges for planetary geophysics and planetary science in general lie with numerical modeling and laboratory studies. For instance, the rheology of planetary matter is not sufficiently well understood as are the properties of materials in the deep interiors of the Earth and Venus, let alone Jupiter and the other giant planets. Numerical modeling has advanced largely in step with the advancement of computer power, but for geophysical modeling of, for example, the dynamo and mantle convection, the achievable parameter ranges are still in many cases far from the realistic ones.

Exploration has started to look beyond the solar system for other planetary systems. About two thousand confirmed extrasolar planets have been discovered to date in the mass and radius ranges from 10 Jupiter masses down to Earth size and even Mars size. Of course, interest is directed towards Earth-like planets in the habitable zone around Sun-like stars in other solar systems! These planets are beyond the reach of present observational tools, and space agencies are planning to launch space-based telescopes that will be more capable such as the planned ESA PLATO mission and the NASA TESS mission. Also, the James Webb Space Telescope will provide unprecedented opportunities.

But the questions are the following: Is the Earth typical? May other solar systems have planets with thick ice covers such as the icy moons of Jupiter, but bigger? Is it conceivable that there are planets of about the Earth's size with primordial, Jupiter-like atmospheres (Stevenson, 2004)? From a theoretical point of view, these worlds are possible (a statement that may simply reflect our ignorance). In the icy planets, radiogenic heat may support oceans and it is conceivable that these oceans harbor life. Could life even exist in gaseous envelopes?

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