Chapter 5: **Interiors of terrestrial planets and major moons**

- basic principles for modeling planetary interiors:
    internal heat; seismology; plate tectonics; magnetic fields
- comparisons of specific objects
- sample problems

Basic principles of planetary interiors

We have a few samples of the interiors of rocky / icy planetary bodies – many meteorites, for example, came our way courtesy of collisions that shattered their original parent asteroids – but by and large we rely on theoretical models supplemented by surface observations. We can observe seismic wave arrival times (at least on Earth), or the rate at which heat is emitted, or where and how frequently volcanoes erupt, or how the walls of craters sag, or how the object's shape deviates from spherical or how the path of a passing spacecraft is affected – all these give us hints about the interior conditions.

The most obvious piece of information about the interior of a planet is its mean density (mass / volume), because that's related to its composition. Roughly speaking, ices have densities of ~1 g/cm^3, surface rocks range from ~2.8 – 3.9 g/cm^3, iron-rich minerals ~4½ – 5½ g/cm^3, and nickel-iron meteorites, ~7 – 8 g/cm^3. Mercury's density is 5.43 g/cm^3. Mercury can't be just rock, but must have some fraction of denser material. Iron is a likely suspect, because it's relatively cosmically abundant. Earth's density is 5.52 g/cm^3. . .but that doesn't mean Earth has relatively more iron than Mercury. The reason is that Earth is massive enough to experience some amount of gravitational compression; our density is higher than the average of the densities of the planet's constituent parts. Understanding the behavior of those constituent parts under the conditions of high pressure that exist at a planet's core would help us disentangle the contributions of compression and composition to the overall density.

We have evidence of grains of dust that pre-date the formation of the solar system; still, near the forming Sun the early solar system was warm and a lot of materials that we now think of as being solid would have been gas. The basic steps in the formation of the rocky / icy planetary bodies include initial condensation of gases into dust grains / snowflakes, accretion of those grains into larger objects, and, for objects that are large enough, differentiation. Refractory materials condense first, while the temperatures in the inner solar nebula are still quite high. At low pressures, elements such as tungsten and zirconium condense at temperatures of ~1,500 K. Oxides of calcium and aluminum soon follow, as evidenced by tiny droplets found in some primitive meteorites (the mineral perovskite, $CaTiO_3$, is one example). Iron-nickel compounds and rocky silicates (e.g., enstatite, $MgSiO_3$) are next, condensing at temperatures closer to 1,000 K. Below 500 K, which mostly means farther out in the solar system than the Earth, it cools enough for materials rich in carbon to condense and for water, ammonia, and methane to form ices. Reality is not quite so neat as these few examples suggest because chemistry keeps happening as dust grains are forming and the material in the protoplanetary disk cools. For instance, iron will react with the gas H_2S at ~700 K, forming troilite (FeS), and will react with H_2O at ~500 K to form iron oxide (FeO). The iron oxide will react with silicates to produce silicate minerals with relatively more iron; olivine, $(Mg,Fe)_2SiO_4$, and pyroxene, $(Mg,Fe)SiO_3$, are two examples. Our most common ices may be water, ammonia (NH_3) and methane (CH_4) but at very cold temperatures, ~25 K, we find CO and N_2 ices instead.

Internal heat

We expect that for smaller objects, e.g. asteroids or moons a few kilometers or a few tens of km in size, the average density and the density of the surface rocks should be more similar because small objects are not large enough to have differentiated. Small objects have less accretional heat to begin with and cool more rapidly because their ratio of surface area to volume is larger. Planets and major moons are large enough to have had enough internal heat to have become at least partially melted, enough for the denser

material to sink into a core the center. The core is surrounded by a less dense mantle and a relatively thin, relatively low-density crust at the surface. Let's look at the sources of heat in more detail.

       Accretional heating refers to the gravitational potential energy released as small planetesimals fall together to form a larger moon or planet. To estimate how much energy will be released requires math that's a tad beyond what we are using, so let's look at it descriptively. Imagine starting with a seed planetesimal to which we are going to add infalling material. The math is easier to handle if we consider a thin shell of material (rather than a rock) falling from infinity onto / around our seed planetesimal. The gravitational energy that will be released on impact is proportional to $-\,G\,M\,m\,/\,r$, where m is the mass of the infalling shell of material and r is the radius of the seed planetesimal where the shell lands. It's negative because we are forming a gravitationally bound system; we'd have to add energy to reverse the process and remove our shell of material. In helping to understand where this value comes from, note that falling from infinity is the precise reverse of being launched with escape velocity. Escape speed is the speed needed, whether by a rock or a rocket, to just escape the gravity of an object, with no extra energy left over. We imagine that a rocket launched with escape speed would have an infinitely long "orbit" with zero speed at that infinitely distant far end of the orbit. Falling from rest at infinity is the reverse of being launched with escape speed. In other words, it makes sense to approximate the speed with which our falling shell will hit the target planetesimal as being equal to the planetesimal's escape speed. (We'll use this same idea in a later chapter when we consider how large a crater an impacting asteroid could make on a planet or moon.) The impacting shell thus has kinetic energy $=\frac{1}{2}\,m\,v^2 = \frac{1}{2}\,m\,G\,M\,/\,r$, substituting in the escape speed squared. It's this kinetic energy that's going into the gravitational energy of the forming protoplanet.

       To get the total gravitational energy of the moon or planet requires summing over all the shells of falling material. Successive shells fall harder and faster, as the mass onto which they are falling, and thus the gravitational attraction, increases. If you've had some calculus, you can see that we need a double integral to get the final gravitational energy (over distance as one shell falls in and over all the shells). Here we'll just state the result, namely that the gravitational energy is proportional to $-\,G\,M^2\,/\,R$. The constant of proportionality depends on how the mass is distributed in the planet.

       The *virial theorem* states that half of the gravitational potential energy goes into the energy of the particles that make up this gravitationally bound system that we just formed. The other half gets radiated away. How hot the interior of the newly formed planetary body will be depends on the composition; i.e., it depends on how easily the material increases in temperature for a given amount of energy added (the *heat capacity)*. Regardless of composition, it should make sense that smaller, asteroid-sized objects aren't going to get as hot inside because incoming material doesn't hit them as hard as it hits, say, the Earth.

       Example: The heat capacity for silicates is ~1000 J / (kg · K) and for the iron and nickel that make up terrestrial planet cores it's ~800 J / (kg · K). If all the accretional energy were produced all at once in an Earth-mass planet that was half rocky and half metallic, and using 3/5 as the proportionality constant in the gravitational energy expression, the temperature rise would be

$$\Delta T = \frac{E}{M \cdot c_P} \approx \frac{\dfrac{1}{2}\dfrac{3}{5}\dfrac{GM^2}{R}}{M \cdot c_P} = \frac{0.3 \cdot 6.67 \cdot 10^{-11}\,\mathrm{m}^3/(\mathrm{kg \cdot s}^2) \cdot 6 \cdot 10^{24}\,\mathrm{kg}}{6.4 \cdot 10^6\,\mathrm{m} \cdot 900\ \mathrm{J}/(\mathrm{kg \cdot K})} \approx 20,000\ \mathrm{K}.$$

Obviously the energy doesn't arrive all at once. Still, more detailed models do indicate enough of a temperature rise to melt at least some of the materials inside a terrestrial planet. Water ice has a larger heat capacity (~2,000 J / (kg · K)) meaning it would be harder to produce a substantial temperature rise in a rock / ice satellite.

The decay of radioactive elements is another significant source of heat for a planetary body. Very early, when planets were forming, radioactive isotopes with short half-lives that were produced in relative abundance in supernovae occurring near the young solar system were important sources of heat. These isotopes include ^{26}Al and ^{60}Fe, with half-lives of $7 \cdot 10^5$ y and $2.6 \cdot 10^6$ y, respectively. Elements with half-lives of billions of years are still producing energy today. Some of the most important today are ^{235}U (half life $7 \cdot 10^8$ y), ^{238}U ($4.5 \cdot 10^9$ y), ^{232}Th ($1.4 \cdot 10^{10}$ y), and ^{40}K ($1.25 \cdot 10^9$ y). None of these elements are extremely abundant (e.g., only ~0.012% of all potassium is ^{40}K) but they are abundant enough to be important factors in the planets' heat budgets.

For moons such as Io, Europa, and probably also Enceladus, variations in the tidal force produce significant amounts of internal heat. Tidal heating depends on several factors, among them the eccentricity of the moon's orbit and the diameter of the satellite. The tidal force is the difference in the gravitational force on opposite sides of the moon; tidal heating results when the tidal force varies with time.

To be complete, there's another source of heating, resistive heating or Ohmic dissipation, that may have been important early in the history of the solar system for conductive objects near the Sun. The magnetic field of the young Sun was much stronger than it is today and the young Sun rotated more rapidly. Moving through that magnetic field could have set up an electric current in an object with a relatively high metal content relatively near the Sun. This isn't a significant factor today but may have mattered for some material near the Sun in the early days of planet formation.

All these sources of heat help, but the larger planetary bodies, with enough internal pressure, don't have to be molten for the rocky material of which they are composed to flow enough for the object to become round (since it's probably rotating, it's likely to become an oblate spheroid) and to differentiate, with the denser materials sinking to the center. If you've ever seen an outcrop of folded rock, you've seen evidence that, given warmth and pressure and time, rock can flow. Models of the formation and differentiation of terrestrial planets suggest that the timescale is on the order of tens of millions of years, i.e., relatively rapid compared to the total lifetime of the solar system.

Figure 5.1: Folded rock.

Chewuch river, northern Cascade range, Washington State.
Photo by AKD.

Hydrostatic equilibrium; phase diagrams
A planetary body that has pulled itself pretty much into round and is stable, i.e., not expanding or contracting, has achieved *hydrostatic equilibrium*. This is described in Chapter 1 (Introduction); the relevant equation is the following:

$$dP = -\rho \, g \, dr,$$

where P is the pressure, ρ the density, g the gravitational acceleration at distance r outward from the center of the object. (You may also see this written as $dP = \rho \, g \, dz$, where z is depth downward into the object.) As expected, this tells us that as we move outward from the relatively higher pressure near the center of a planetary body the density will drop.

Example: the density and gravitational acceleration inside the Earth are clearly not constant but if we made the approximation that ρ and *g* were constant we could get a (very) rough first estimate for the Earth's central pressure. Using the average density, the surface gravitational acceleration, and the average radius of the Earth, we have

$P_{\text{core, estimated}}$ = 5,500 kg/m³ · 9.8 m/s² · 6.37·10⁶ m = 3.4·10¹¹ Pa (or 3.4 Mbar).

That's an underestimate, but only by ~6%.

A note on units: the SI unit of pressure is the pascal; 1 Pa = 1 N/m² = 1 kg / (m s²). We often still use the bar, with 1 bar = 10⁵ Pa, and the atm (or standard atmosphere), roughly a measure of sea level air pressure at mid latitudes; 1 atm ~101 kPa. You may also see barometers that read in mm of Hg (which are approximately equal to Torr) or measure tire pressure in psi (pounds per square inch); 1 atm ~760 mm Hg ~ 14 psi. Too many units!? Agreed.

The equation of hydrostatic equilibrium does not tell us whether the density changes smoothly or discontinuously with radial position inside a planetary body. To describe the state of the material at some location inside a planetary body more fully we need an *equation of state*, meaning a way to relate the pressure, density, and the temperature of the material. Gases can often be treated as if they obey the ideal gas law but rocks are more complicated. One reason is that with changes in pressure and/or temperature substances can change their state or phase. The change you are likely most familiar with is melting from a solid to a liquid, but solids themselves may change state. Carbon provides a good example: graphite is soft and, mixed with clay, is commonly used to make pencil lead (which doesn't have anything to do with the element lead); carbon atoms structured instead in a rigid lattice gives diamond, the hardest natural substance. Diamond and graphite are two states of solid carbon.

Solid water, as another example, has quite a few phases. Some are amorphous, some crystalline; ice Ih (the "h" stands for hexagonal) is the only phase you are likely to encounter on Earth, but moderately high-pressure phases likely play a role in the interiors of icy outer solar system objects (and supercritical fluids in the giant planets). A very detailed, often-quoted, phase diagram for water may be found at http://www1.lsbu.ac.uk/water/water_phase_diagram.html. The following, less-detailed, sketch is based in large part on that diagram. The locations indicated by giant planet interiors and Ganymede's (and similarly sized ice-rock moons) interior are approximate.
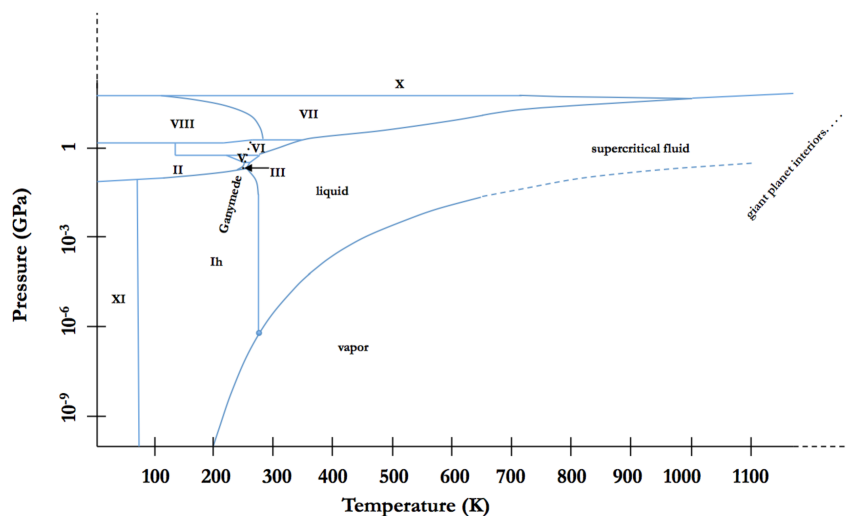

Figure 5.2: Phase diagram, water

In 2018 and 2019 researchers reported creating 'superionic' ice XVIII at higher temperatures and pressures (above 5,000 K, ~ 40 GPa) than shown on the phase diagram above. Strictly speaking, this isn't exactly water any longer, because under these conditions hydrogen atoms separate and are able to wander through an oxygen matrix, behaving a bit like electrons in a metal. Ice XVIII is about 4 times as dense as normal ice and curiously black in color. It's possible, given the expected temperatures and pressures in the 'mantle' layers of Uranus and Neptune, that superionic ice might play a role in the dynamo generating the magnetic fields of the slush giants.

There are many places where three phases meet; what's customarily meant by the *triple point* is the pressure – temperature combination where a substance can exist as solid, liquid, and gas. We will return to this concept when we are considering the surfaces and atmospheres of rocky / icy planetary bodies – conditions on Earth are near the triple point for water, while conditions on the surface of Titan are near the triple point for methane.

The existence of different phases means that the interior of a planetary body could be complicated physically, involving multiple layers, even if it were quite similar chemically. As one example, the Earth's outer core is liquid (which we know from seismology, described below) while the inner core is solid. As another, in the Earth's mantle, at a depth of ~400 km, the density jumps by ~10% due to the transition of olivine to denser spinel. The temperatures and pressures at significant depths are hard to test by replicating them in the lab. Using a diamond anvil press it's possible to sustain pressures of at least 1 Mbar (100 GPa), but not necessarily in conjunction with sufficiently high temperatures. Mantle xenoliths provide another window into conditions below Earth's crust. A xenolith is a rock fragment that gets embedded or caught up in a larger rock. Some xenoliths are fragments of upper mantle rock that get picked up by rising magma and thus delivered to the surface of Earth. If they rise to the surface rapidly enough they will retain a record of the mantle conditions under which they formed.

One place we definitely expect to find density variations is near the surface of a planetary body. An ice cube floating in a glass of water is in equilibrium, partly above and partly below the surface of the water. The ice floats because it is less dense than liquid water; most rocks don't float. How far the ice cube sinks into the water also depends on the water; for instance, salty water is more buoyant. Various features rise up above the mean surface level of planetary bodies, e.g., volcanoes, crater rims, sky scrapers. If the planetary body is in hydrostatic equilibrium we expect to find *isostatic compensation*, in which the layers under the mountain compress a bit to balance the weight of the mountain floating, like the ice cube, partly above the surface. Here's a sketch to show what happens on Earth. Continental crust is less dense than ocean crust (roughly, it's the difference between granite, ~2.7 g/cm^3, and basalt, ~3.0 g/cm^3), both of which are less dense than the upper mantle on which they ride. Like the ice cube, the crust sinks a bit into the mantle.
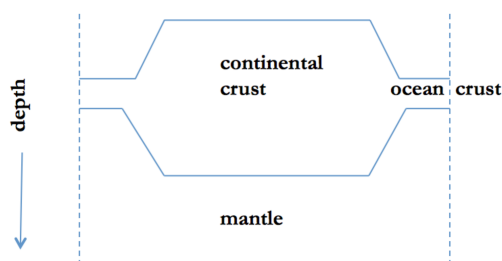

Figure 5.3: Isostatic compensation

The icy surface of Europa shows much less vertical relief than our Moon does. Images of the surface of Pluto from the recent New Horizons flyby show icy mountain ranges, at least one of which is as high as the Rockies. It's clear that Pluto's crust is able to support taller mountains than Europa's can and thus Pluto's crust must be stiffer than Europa's. Given how cold it is, that is no shock!; on the other hand, some models of Pluto's surface features suggest that Pluto might also have a subsurface layer of liquid. That hypothesis

is supported by the presence of ammonia near cracks on Pluto's surface; ammonia should be destroyed by solar UV, meaning if it's on the surface now it would need to have been replenished, in something on the order of a few hundred million years, perhaps from an underground reservoir of liquid.

Over time every planetary body will cool, regardless of how hot it may have been early in its history. Cooling involves the transfer of heat from the interior to the surface and the radiation of energy into space. As already noted, smaller objects, with a larger surface area-to-volume ratio, are going to cool faster. Convection is very efficient at moving energy outward, sewing the seeds of its own demise; convection ceases once the interior of a planetary body is too cold and stiff for anything to move. Conduction varies in its efficiency; obviously planetary bodies made of materials with a higher thermal conductivity will cool faster. Heat provided by radioactive elements will graduate die out. Keep in mind that cooling can involve phase changes – think of water freezing into ice. As a practical result, we would expect, for instance, that the Moon should be relatively more solid inside than the Earth and that the crust of Mars is thicker now than it once was.

Seismology

Seismology, the study of the propagation of waves through the Earth, has given us the best insight into the conditions in the interior of the Earth. We also have several years' worth of data from the Moon, from seismometers left by several of the Apollo lunar landing missions (and may in the not-too-distant future include seismometers on Mars missions).

There are three categories of seismic waves: body waves, which travel through the interior, surface waves, which travel along the surface, and standing waves or free oscillations, usually following very strong earthquakes, in which the planet rings like a bell. (The Sun exhibits free oscillations all the time and analysis of the surface manifestations of these waves by helioseismologists has provided our best insights into the state of the interior of the Sun.)

Body waves can be either longitudinal or transverse. The longitudinal waves are called P waves, for pressure or primary. Like sound waves, these waves propagate by the systematic compression and rarefaction of material in the direction the wave is traveling. Pressure is the restoring force. Both solids and liquids will support the propagation of P waves. That's not the case for the transverse S waves. In these shear waves material moves back and forth perpendicular to the direction the wave is propagating, similar to electromagnetic waves; that works in solids but not in liquids. The S waves are also called secondary waves, because in solids P waves travel faster and arrive at seismometers sooner than S waves. The U.S. Geological Survey has helpful sketches to aid in visualizing these waves (http:// earthquake.usgs.gov/learn/glossary/). Imagine hitting a column of rock with a hammer. Hit it on the end and a pressure wave will propagate along the column; hit it on the side and the column will ripple back and forth. There are also two types of surface waves. Rayleigh waves produce an elliptical ground motion that's up and down (similar to water waves) not side-to-side, and Love waves produce horizontal motion.
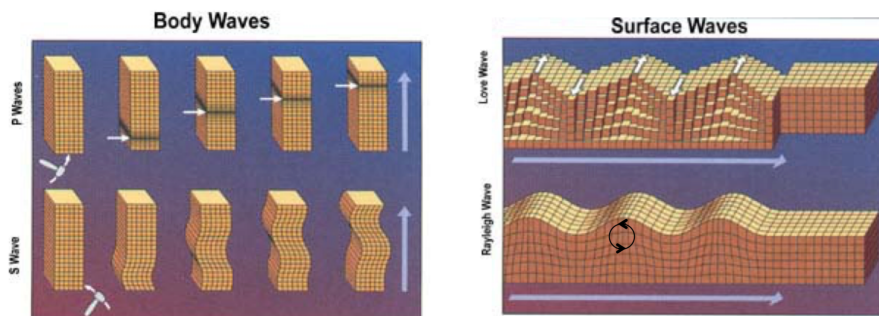

Figure 5.4: Seismic waves   http://earthquake.usgs.gov/learn/glossary/

The focus (or hypocenter) is the local of the earthquake; the epicenter is the location on the surface directly above the focus. The velocities of the P and S waves depend on the properties of the medium through which they are travelling – e.g., the density, the rigidity, the compressibility. Both velocities increase with increasing density. Velocities of seismic waves in the Earth range roughly from 3 – 13 km/sec, with P waves travelling about 1.7 times faster than S waves. Densities change along a wave's path and, just like with electromagnetic waves, the seismic waves will experience refraction. The path of a wave will be curved upward, gradually, where the density is increasing smoothly, and with some discontinuities where the density changes abruptly.
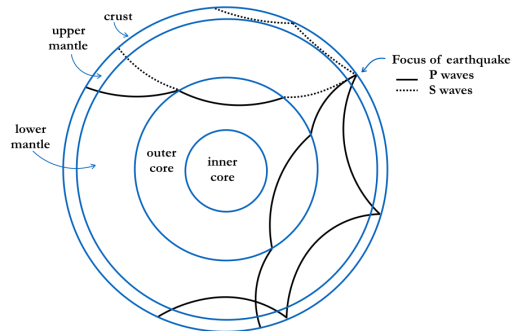

Figure 5.5: Waves can also be reflected from the surface, as shown in this sketch.

Waves, particularly from large earthquakes, can travel all the way across the Earth; shear waves can't travel through the liquid outer core but they may be converted to compressional waves at the core / mantle boundary. The arrival times of the P and S waves are noted at many seismic stations, from which data the location of the focus, the paths of the waves, and thus the properties of the Earth's interior, can be constructed. We've already noted that S waves are blocked by the liquid outer core. Because of the refraction note that there are also "shadow zones" that won't see P waves. Specifically, seismometers located between 104° and 140° from the epicenter won't record P waves.
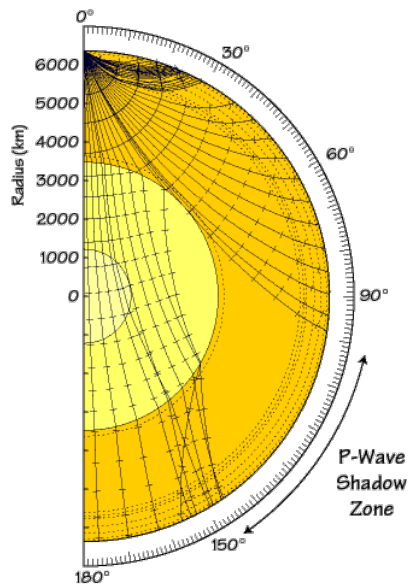

Figure 5.6: P waves from a shallow quake propagating through the Earth.

(This source web site, as of August 2015, has a flash animation of waves propagating through the Earth.)
http://earthquake.usgs.gov/learn/glossary/?termID=170&alpha=S

The following sketch shows schematically what a seismogram (i.e., what's recorded by a seismometer) generally looks like.

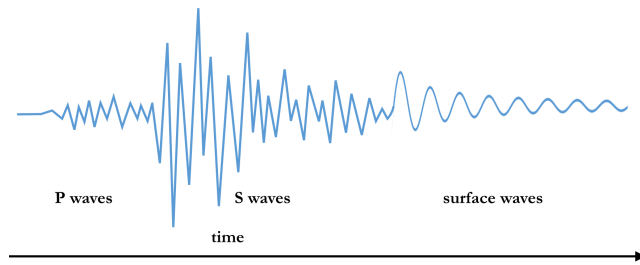P waves   S waves   surface waves

time

Figure 5.7: example seismometer recording

Plate Tectonics

If you look at a world globe it's easy to see that Africa and South America look as though they should fit together, like pieces in a jigsaw puzzle. Accepting that they might have once fit together, that continents could "drift", was not easy. In the 1950s and 60s geoscientists were able to identify mid-ocean spreading ridges, where new crust is being formed, and deep trenches where dense old crust is sinking, or being subducted, under less dense crust; these discoveries helped make the new idea of plate tectonics more acceptable. Today we would say that our *lithosphere*, the solid crust and outer mantle of Earth, is broken into multiple plates that float on a less rigid mantle layer called the *asthenosphere*. Huge slow convective cells, helped along by the push of spreading ridges and the tug of slabs sinking at subduction zones, move the plates at rates up to about 0.1 meter per year. The following sketch illustrates this basic idea:
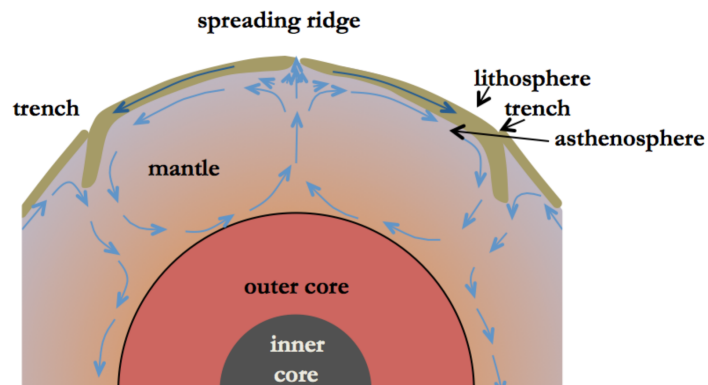

Figure 5.8: Plate tectonics

The new crust formed at spreading ridges is typically basalt, which is somewhat denser than rocks such as granites that make up much of our continental crust. It cools and thickens as it gets carried away from the warm spreading ridge. Near the ridge the lithosphere might be barely 10 km thick; across a large ocean, about to dive into a trench, the ocean lithosphere might be more than 100 km thick. Continental lithosphere, which includes the less dense continental crust, is thicker, ~200 km.
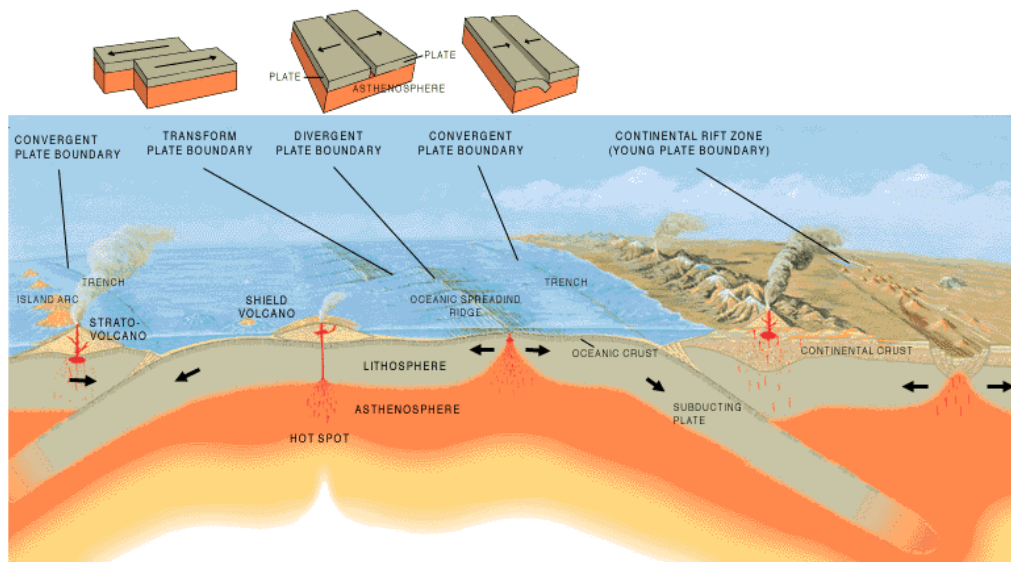
Figure 5.9: Plate tectonics, a bit more detailed; http://pubs.usgs.gov/gip/dynamic/Vigil.html
Illustration by José F. Vigil in *This Dynamic Planet*, USGS / Smithsonian Inst. / US Naval
Research Lab.

     Plate boundaries are geologically active locations. It's along plate boundaries that we find many of the Earth's active volcanoes and the majority of noticeable earthquakes. The above illustration from the United States Geological Survey (USGS) shows the variety of events that occur along plate boundaries. The inset at the top of this figure shows the ways plates could meet. They could slip or jerk past each other at a transform boundary; this is what's happening along California's San Andreas fault, where the Pacific Plate is moving northward with respect to the North American Plate. A divergent boundary is the general name for plates spreading apart. A mid-ocean ridge is one type of divergent boundary, and the Mid-Atlantic ridge, shown in the next illustration, is a good example. Note that Iceland sits atop the Mid-Atlantic ridge and spreads by a few centimeters per year. A continent can also split; Africa has been doing so along the East African Rift zone for the past 20+ million years, and this ever-increasing split will likely give rise to a new ocean basin in another 10 million years or so.
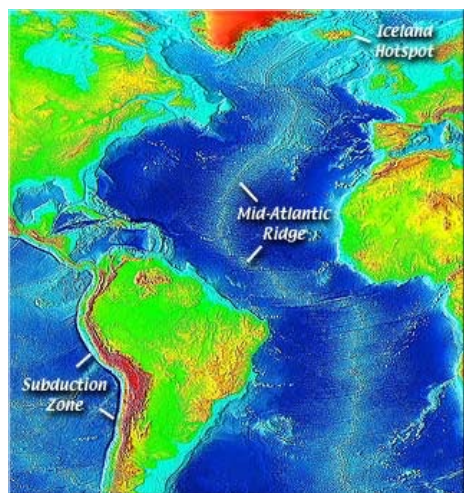

Figure 5.10: Mid-Atlantic Ridge

http://geomaps.wr.usgs.gov/parks/province/atlantpl.html

     Convergent boundaries are where plates collide. One type (not shown in the illustration above) of convergence occurs when two continents collide. A high level of crumpling ensues, resulting in mountain

ranges such as the Alps and Himalayas. Dense ocean lithosphere colliding with a continent results in a subduction zone, where the denser plate sinks beneath the lighter continent. Material that has been eroded by rain and rivers collects in the ocean along the continental margins. A subducting slab will scrape up some of this sediment and carry it along down into the mantle. Two plates could also collide where ocean crust meets ocean crust; one will usually be denser and sink. At a subduction zone a deep ocean trench often forms, such as the 11-km-deep Marianas trench in the western Pacific. The sinking slab and its attendant sediment, particularly the hydrous minerals, will partially melt, contributing magma to chain of volcanoes ~2-300 km from the trench. The Andes mountains in South America and the Cascade range in the Pacific Northwest are examples of mountains built along the leading edge of a continent near a subduction zone. Magma doesn't only rise to the surface along spreading ridges or behind subduction zones. The sketch above includes a mantle hot spot; hot spots give rise to volcanoes such as the Hawaiian islands or Yellowstone. Here is a map showing the major terrestrial tectonic plates:
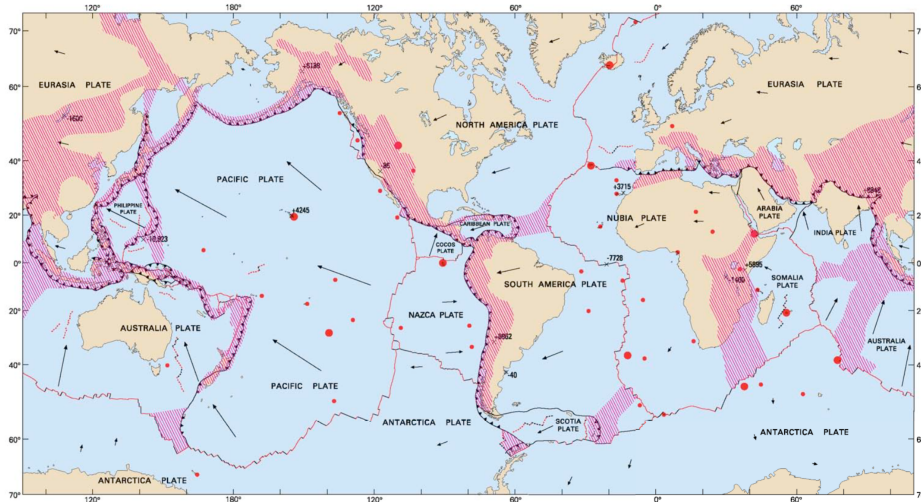

Figure 5.11: Tectonic plates

http://pubs.usgs.gov/imap/2800/TDPfront-screen.pdf

The Pacific Ocean is bordered by plate boundaries. That border is often called the "ring of fire" because there are so many earthquakes and volcanoes around the rim of the Pacific. The following image shows the locations of earthquakes (black) since ~1750 and volcanoes (red) over approximately the last 10,000 years.
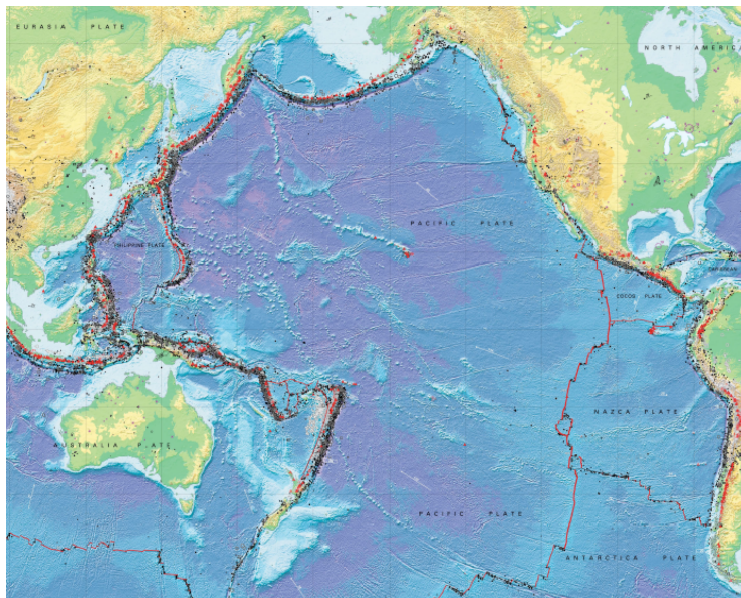
Figure 5.12: Ring of Fire

from
This Dynamic Planet
Smithsonian Institution, USGS,
US Naval Research Laboratory
http://pubs.usgs.gov/imap/2800/TDPfront-screen.pdf

Notice also in the above figure that the Hawaiian island chain extends well beyond the group of major islands that make up the State of Hawaii. Here we have an example of a volcanic hot spot. The Pacific plate has moved across the top of the hot spot, producing a chain of volcanoes. The oldest, most eroded, members of that chain lie at the northwest end. (As it gets farther from the hot spot the oceanic basalt will cool and subside a bit, contributing some sinking to the erosion of the northwest end of the Hawaiian chain.) The youngest is the Lōʻihi seamount, about 35 km SE of the island of Hawaii and still nearly a kilometer below sea level.

Does plate tectonics occur, or has it taken place in the past, on other solar system bodies? Maybe. Tectonic activity, in the form of faults or rift valleys, is clearly present on many terrestrial objects but there is no definitive evidence for the current presence of moving plates on anything other than Earth. There are some indications that plate tectonics might have gotten started on Mars but that the crust thickened too quickly for sustained plate motion. One reason Olympus Mons is so large is that, unlike the Hawaiian chain of volcanoes, the hot spot that created Olympus Mons kept making a volcano in one location. The Pacific Plate, moving across the Hawaiian hotspot, made a volcanic chain; Olympus Mons, sitting atop its hot spot with no plate motion, just got bigger. Another piece of evidence is that features that may be ancient river beds on Mars don't show evidence of having been shifted by the topographic upheaval that would accompany plate motions. Features on Titan that appear to be rivers flowing with liquid methane resemble Martian structures more than they resemble terrestrial rivers, suggesting that, like Mars, Titan also doesn't have active plate tectonics.

There have been suggestions that we see something akin to plate tectonics taking place on Europa. Blocks of ice on Europa show strong evidence of moving, splitting, rotating, etc., atop a liquid layer. Whether they subduct is debatable. The following sketch illustrates features that indicate the possible presence of a subducting plate. It's not necessarily clear why one icy plate would be dense enough to sink, since water ice is less dense than liquid water, although it could possibly be driven down by crustal extension occurring upstream at a spreading ridge.
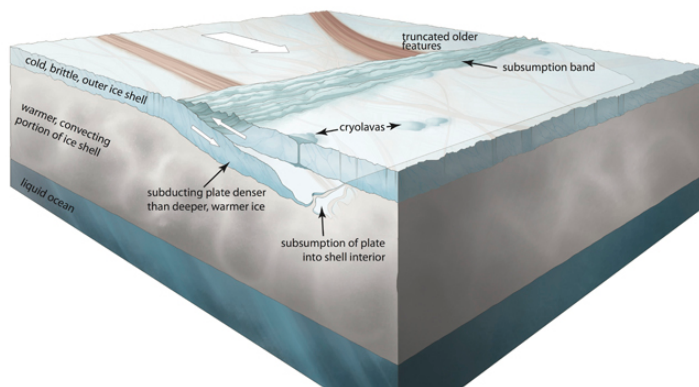
Figure 5.13: Model of an icy subducting plate on Europa.

Image: Noah Kroese

Magnetic fields

Most planets are surrounded by a magnetosphere, a region within which the planet's magnetic field dominates over the Sun's interplanetary magnetic field. This bubble can be more than 100 times the size of the planet; magnetospheres are described briefly in the planetary overview chapter. Some planetary bodies, such as our Moon, although not actively generating magnetic fields, still have remanent fields. Some have temporary fields by virtue of interacting with the solar wind particles or the field of their local giant planet. Some planetary bodies are still actively generating magnetic fields in their interiors.

Planets aren't simply permanent magnets – iron, for instance, loses its magnetism at temperatures greater than ~1,000 K. At lower temperature iron atoms tend to line up (their dipole moments align) but as the temperature increases, the atoms' kinetic energies prevent them from lining up and creating a magnetic field. The temperature at which materials lose their magnetism is called the Curie temperature (named for Pierre Curie, husband of Marie and also a physicist). Even at cooler temperatures permanent magnetism eventually dissipates; for Earth, the dissipation time has been calculated to be much less than a million years. The magnetic field must be continually being regenerated.

We are sneaking up on *magnetohydrodynamics*: the (re)generation of magnetic fields by the motion of electrically conducting fluids. What follows is a simplified description of how this works. We are going to make use of the principle that moving electric charges induce magnetic fields and that changing magnetic fields induce electric currents. Start with an initial dipole (N/S) magnetic field, such as is shown in the following image of iron filings around a bar magnet, and a liquid conducting layer in a
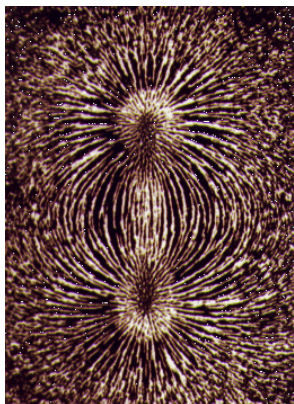

Figure 5.14: Dipole magnetic field
http://solarscience.msfc.nasa.gov/magmore.shtml

rotating planet with an internal heat source. Introduce convection: bubbles of warmer, more buoyant material will rise. If this material is conductive and dense enough, the magnetic field lines will be shoved around by the material (rather than the other way around, as in the familiar examples where magnetic field lines drag around a compass needle or nearby iron filings). Because the planet is rotating, the field lines will become twisted as they rise with the convective bubbles. This introduces a toroidal component to the magnetic field (i.e., an "around" component in addition to the "up-down" dipole component). That toroidal component gets twisted and induces more dipole field which gets twisted. . .this process is called a magnetic dynamo.

The dynamo is the best model for sustaining magnetic fields such as those in the Sun, giant planets, Ganymede, and Earth. It's unclear whether Mercury has a magnetic dynamo, although recent reports of the extent to which the crust of Mercury seems to "float" rather than rotate in lockstep with the planet's core suggest a liquid layer, around a solid inner core, and support the existence of a dynamo. The field on Mars is likely remanent magnetism from a dynamo in the distant past. Europa and Callisto seem to have magnetic fields that are induced by their motion through Jupiter's magnetic field, rather than having dynamos of their own.

One aspect of the magnetic dynamo that provides some insight into the history of the Earth is that the re-induced dipole field is oppositely oriented; there has been a *magnetic reversal*. In other words, if we start with a magnetic north pole at or near the geographic north pole, after some time the pole will flip and magnetic north will be at or near the geographic south pole. Later it will flip again, and again, and yet again. In the Sun, we observe that during a given sunspot maximum the orientation of the magnetic pole is opposite that of the preceding sunspot cycle. The Sun's pole flips, on average, about every 11 years. Earth's pole flips every few hundred thousand years. On average, it's about 450,000 years between flips; the reversals themselves may take a few thousand years. There's a huge standard deviation on that 450,000 year average, but still, we do seem to be overdue for a reversal – the last one was ~780,000 years ago.

The orientation of the Earth's magnetic field is recorded in iron-rich magmas. Magmas erupt at temperatures above the Curie temperature for common iron-rich minerals. As the lava cools, the magnetic "domains" align with the prevailing direction of the magnetic field. Once the temperature drops enough, the magnetic alignment is frozen in to the rock. One of the first convincing pieces of evidence supporting the idea that the Mid-Atlantic Ridge is a spreading center was provided by maps of magnetic "stripes" on the Atlantic basin. Lava recently erupted on the ocean floor would show a magnetic field aligned with the current orientation of Earth's magnetic field. Lava erupted 800,000 years ago, and having been carried away from the ridge, would be oppositely aligned. The following sketch illustrates the pattern that develops on the seafloor to the sides of a spreading ridge as time progresses from *a* to *b* to *c*; the model times shown are 5 million years ago, 2.5 million years, and present day.
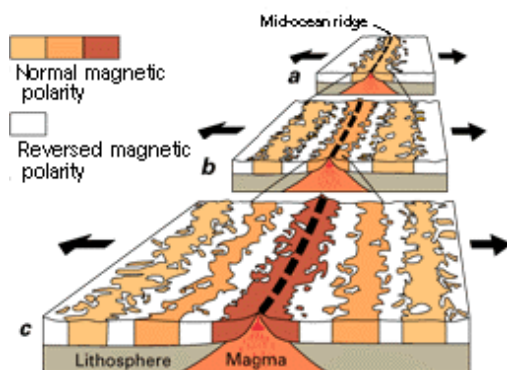

Figure 5.15: Magnetic field orientation record at a mid-ocean ridge

http://pubs.usgs.gov/gip/dynamic/developing.html

Here too one might ask whether we see similar magnetic field patterns anywhere else in the solar system indicative of magnetic field reversals. Again, the answer is "maybe". There are some hints of magnetic striping on Mars. It's not recent, but suggests that the early Mars might have had incipient plate tectonics, a dynamo, and magnetic field reversals that were unsustainable once the lithosphere thickened too much.

Comparisons of specific objects

The following plot of density as a function of depth inside the Earth today is from the Preliminary Reference Earth Model, developed by Dziewonski and Anderson in the early 1980s. There are some distinct discontinuities within the mantle, not just between the lower mantle and outer core. At depths of ~410 km and ~670 km seismic wave velocities increase, indicating an increase in density.
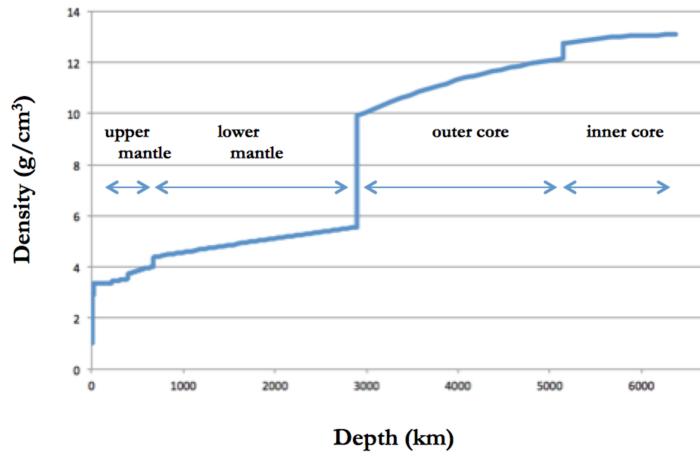

Figure 5.16: Preliminary Reference Earth Model
http://ds.iris.edu/spud/earthmodel

Our current model is that these depths correspond to pressures where olivine, a major mantle constituent, undergoes a phase change to a denser crystal configuration.

For most rocky planetary bodies density, rotation, and gravitational and magnetic fields permit us to construct general interior models even if we lack the seismic observations needed to create a detailed model. Without seismic data, though, estimates of size of the core involve a bit of guesswork about composition. For example, a core made of iron and nickel could have a radius half the size of a core made of iron and nickel plus a substantial percentage of (less dense) sulfur. The following sketch shows the (approximate except for Earth) relative sizes of the metallic cores and rocky mantles for the major rocky planetary bodies. Two points to note: Mercury's core is relatively large and the Moon's core is relatively small. Mercury must have relatively more iron than other inner planets and the Moon must have relatively less iron.
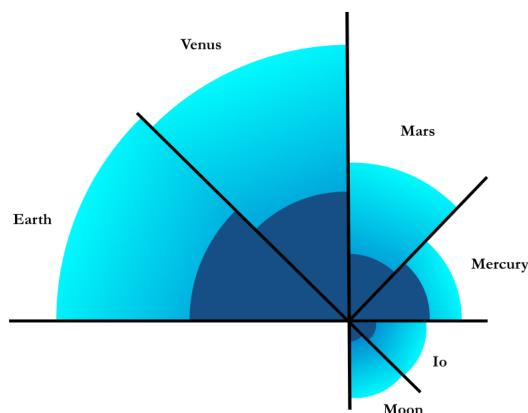

Figure 5.17: Core sizes for rocky objects.

Modeling the large icy outer solar system objects is more complicated. Water ice, as noted above, has many different phases. Water with additives such as salts and organic molecules has a lower freezing point than pure water; this plus a sufficient heat source makes subsurface oceans possible on objects such as Enceladus, Europa, Titan, Ganymede, Callisto, maybe Ceres, possibly even Pluto, and perhaps others. Tidal heating and a salty ocean works well for Europa. Tidal heating alone would seem to be an

insufficient heat source for Enceladus; one possibility is a core that's not solid but more like gravel and thus subject to more friction. Continuing decay of radioactive isotopes in the interiors of some otherwise cold objects might be enough to maintain an internal liquid layer today.

Ceres has been observed both from the ground and from the Dawn spacecraft, in orbit around Ceres since early 2015. Despite being less than 1,000 km diameter, Ceres is slighly oblate, consistent with its being in hydrostatic equilibrium and at least partially differentiated. It's likely to have an ice-rich mantle, although whether that includes a liquid layer is unclear.

One recently proposed model for the interior of Ganymede posits the existence of multiple layers of alternating ice and liquid around a silicate mantle and metallic core, as shown in the following drawing:
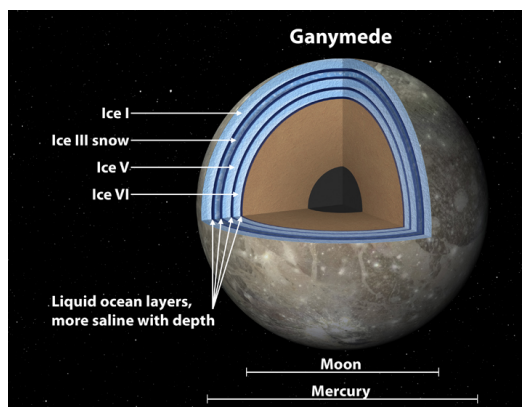
Figure 5.18: Ganymede interior model

NASA / JPL
https://photojournal.jpl.nasa.gov/catalog/PIA18005

In general it is not yet possible to make definitive statements about the thickness of icy crusts, depth of ocean layers, core size, and so forth for the moderate-to-large-size icy planetary bodies.

Table 5.1: **Densities and magnetic fields of selected terrestrial solar system objects**

object	average density (g/cm³)	surface magnetic field (µT)	field source / comments	interior liquid layer?
Mercury	5.43	0.25	dynamo?	partially molten core?
Venus	5.24	$\sim 5 \cdot 10^{-4}$	induced by solar wind	Fe-Ni core?
Earth	5.51	31	dynamo	Fe-Ni outer core
Moon	3.35	~0.03	localized crustal field	thin outer core
Mars	3.93	$\sim 10^{-3}$	loc. crustal field + solar wind	Fe/FeS core?
Io	3.53	~1.3	interacts with Jupiter's field	thin outer mantle shell
Europa	3.01	~0.2	interacts with Jupiter's field	ocean
Ganymede	1.94	~0.75	intrinsic; dynamo?	ocean; possibly two
Callisto	1.83	~0.04		ocean? not completely differentiated
Titan	1.88	~0	interacts with Saturn's field	ocean
Enceladus	1.61	~0	interacts with Saturn's field	ocean
Pluto	1.86	no?		ocean?

Models suggest subsurface oceans are possible on Rhea, Titania & Oberon, Triton, Orcus (a plutino), Eris, Sedna (TNO; very long eccentric orbit)

Sample problems

1. Estimate the radius of the Earth's core relative to the total radius if the
   core density $\approx$ 11,000 kg/m³,
   mantle density $\approx$ 4,500 kg/m³,
   average density $\approx$ 5,500 kg/m³

2. The equation for hydrostatic equilibrium is $\Delta P / \Delta r = -\rho g$. Crustal rocks have densities of ~ 3 g/cm³ and for the Earth surface $g = 9.8$ m / s². Estimate the pressure at 5 km depth under the surface of a terrestrial continent; express your answer in GPa.

3. Olympus Mons is huge and the Hawaiian islands are a chain. Describe our current model for explaining this difference.

4. Explain how we can use magnetic field orientations at mid-ocean spreading ridges to determine that the Earth's magnetic pole flips.

5. Explain the evidence that suggests that part of the Earth's core is liquid.

6. The tidal force matters for tidal heating. Calculate the ratio of the tidal force on Io due to Jupiter to that on Enceladus due to Saturn.

7. Reading carefully? Briefly explain or define
   a) accretional heating
   b) triple point
   c) lithosphere
   d) asthenosphere
   e) ring of fire
   f) isostatic compensation
   g) why we expect that a layer of liquid water under the crust of an outer solar system satellite would likely be salty

Answers to selected questions are on the next page.

1. $r_{core} / r_{total} \sim 0.54$

2. Pressure at 5 km depth ~ 0.15 GPa

6. ~3600