Planet Earth

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"It is my opinion that the Earth is very noble and admirable... and if it had contained an immense globe of crystal, wherein nothing had ever changed, I should have esteemed it a wretched lump and of no benefit to the Universe."

Galileo

The study of the outsides of planets has been tremendously rewarding to planetary scientists. The surface of a planet provides clues as to what is happening inside but it is also shaped by impacts and erosion and many of the clues are ambiguous. For example, mankind has been studying the surface of the Earth since he arrived on it but it has only been in the past several decades that he has appreciated the internal turmoil that continuously shapes the surface. Man was learning about continental drift and plate tectonics on the Earth at the same time he was learning that each other world has a unique style of operation. As far as we know the Earth is the only planet that has active plate tectonics, oceans and life and one wonders if these facts are related.

Geophysicists study the interiors of planets. The Earth is the only body for which we have detailed information, including three-dimensional images of the internal structure, from the surface to the center. Even though planets do not all behave the same today they may have been put together in similar ways and intensive study of the Earth's interior may yield knowledge that is transferrable to other planets. We will start, however, by showing how our views of the Earth have been shaped by the study of other worlds.

Comparative Planetology

The Earth is the largest terrestrial planet and contains slightly more than 50% of the mass in the inner Solar system, excluding the Sun. Most of the surface of the Earth is less than 100 million years old, and even the oldest rocks are less than 4000 million years old, so the record of the origin of our planet has been erased many times. Part of this erasure is due to wind and water erosion and part is due to the continual recycling of material back into the interior and the repaving of the ocean basins by sea floor spreading.

The other solid planets and smaller worlds have much more ancient surfaces and this tells us two things that we could not learn from the Earth itself. One is that violent and gigantic impacts were common in the early days of the Solar system as the larger bodies swept up and devoured the smaller objects. The other thing is that most other worlds preserved evidence of these early happenings and the Earth did not.

Earlier views of Earth origin envisioned a gentle rain of dust and small particles which slowly built up the Earth layer by layer. A slowly accreting planet remains relatively cool and it builds up heat mainly by the slow decay of radioactive elements. In this scenerio a cold, homogeneous Earth eventually heated up, started to melt and formed a buoyant crust and a dense core, somehow leaving behind, in most versions of this story, a homogeneous mantle.

However, the energy associated with a large impact is enough to melt, or even vaporize, a large part of both the impactor and the planet. The currently popular scenerio for forming the Moon by impacting the Earth with a Mars sized object serves to melt a large part of the Earth. Even smaller hits, of which there were many more, would melt the portion of the Earth into which they penetrated, and sent shock waves.

The original controversy about whether the Earth started out hot or cold depended on whether it accreted slowly, 10^8 or 10^9 years, or rapidly 10^6 - 10^7 years. In the later case the gravitational energy is being delivered faster than it can be conducted and radiated away and the planet is at the melting point, at least in the outer parts, as it accretes. Every giant impact, however, is essentially an instantaneous accretion event and planets growing by gathering up relatively large objects will experience widespread melting, over and over again. This means that planets, as they grow, have the opportunity to separate their materials depending on melting points and densities. The dense materials will sink toward the interior and the light meltŝ and crystals will rise to the surface. Such a process is called differentiation or gravitational separation and such a planet is gravitationally stratified by a zone-refining type process. The light crust and the dense core are two of the more obvious products of this zone-refining process. The mantle itself can be layered according to chemistry and density.

A magma ocean has been used to explain the anorthositic highlands (buoyant crystals that floated to the surface), and trace-element enriched (KREEP) basalts (the final liquid dregs of a crystallizing magma ocean) on the Moon. As the magma ocean cools and crystallizes the various crystals and residual melts stratify themselves according to density. A similar process probably occurred on the Earth, except the pressures are much higher and dense garnet bearing eclogite forms instead of buoyant anorthosite. The magma ocean concept would not have been thought of by terrestrial geologists because one major product of the ocean sank instead of floated, and thereby disappeared from view.

The spin axis of a planet is controlled by the distribution of masses on the surface and in the interior. If a large impact or a new volcano rearranges this mass distribution then the planet will reorient itself relative to the spin axis so that the mass excesses are close to the equator. The equator of a rotating planet is, in fact, defined by the integrated effect of the mass anomalies. The motion of the pole relative to the planet is called "true polar wander". Both Mars and the Moon have apparently reoriented themselves to accommodate mass redistributions caused by impacts or volcanoes. The spin axis of a spinning top is controlled by its shape, and it will change if bits of clay are attached to the surface. The physics of planetary reorientation is the same.

On the Earth true polar wander is occurring today mainly as the result of mass rearrangements due to deglaciation. Usually, however, the evidence for ancient polar motion found in the magnetic crystals of rocks is treated as "apparent polar wander" since we know that the continents have been drifting relative to the magnetic pole. Our extraterrestrial experience, however, now tells us that periods of a major readjustment of the Earth's shell relative to the spin axis might be expected, the result of mass readjustments in the interior due to convection, subduction or continental insulation, and these might be responsible for some major events in the geological record, such as the breakup of supercontinents.

These planetary lessons are important to students of the Earth. But the Earth has many more lessons for the students of other planets and the Solar system.

The Earth's Interior

The crust, mantle and core are the major regions of the Earth's interior and they account for, respectively, 0.4%, 77.1% and 32.5% of the mass of the planet. This division was initially based on physical properties such as density and seismic velocity rather than chemistry or composition. All of the physical properties vary with depth because of the increase of temperature and pressure and, in places, changes in chemistry or physical state. For example, there is a large jump in seismic velocity at the crust-mantle boundary (the "Moho" discontinuity) because the chemistry changes and there is a large decrease in the compressional wave velocity at the core-mantle boundary because the composition changes from silicate to metallic and the physical state changes from solid to liquid.

Smaller discontinuities occur near 400 and 650 km in the mantle and also near the bottom of the mantle and half way through the core. These are often attributed to phase changes (an example of a phase change is displayed by carbon which is graphite at low pressure and diamond at high pressure) but they can also represent changes in chemistry.

There are also large variations in seismic velocity from place to place at the same depth. These "lateral variations" have been revealed by seismic tomography. In the upper mantle the Earth has as much variation horizontally as it does vertically.

The most common crustal minerals (plagioclase, quartz) and mantle minerals (olivine, pyroxene, garnet) are unstable at high-pressure and they exist only in the crust and the shallow mantle. As pressure increases the atoms in the crystals become more tightly packed and the density and seismic velocities increase. These changes are gradual except at phase changes, or changes in crystal structure, where they can occur over a small depth or pressure interval. This causes a rapid or abrupt change in the physical properties including those measured by seismic techniques. When the change occurs abruptly it is called a seismic discontinuity. The two largest of these in the mantle are at 400 and 650 km and these represent discontinuous rearrangements of the atoms in the major mantle minerals.

Seismic discontinuities allow a further division of the Earth into crust, upper mantle, lower mantle, outer core and inner core. These regions are not necessarily all chemically distinct nor can we assume that each of these regions is chemically homogeneous. The region between the 400 and 650 km discontinuities is called the transition region or the mesosphere (for middle mantle). It has been suggested that midocean ridge basalts and new oceanic crust is derived from this region.

Just above the core is a 200 or 300 km thick layer, called D", that may differ chemically from the rest of the lower mantle. This layer may represent material that was once dissolved in the core or dense material that sank through the mantle but was unable to sink into the core. D" is about 6%, by mass of the Earth or about 9% of the mantle.

Composition of the Earth

The Earth is the only planet for which we can speak with some confidence about the bulk composition, or chemistry, of the whole planet. From the mass of the Earth and seismic determinations of the radius and density of the core we know that the Earth is about one-third iron and that it is concentrated toward the center of the planet. Seismology also tells us that the outer part of the core is liquid, a strong indication that it is molten and behaves in general like other fluid parts of the Earth. The small solid inner core, less than the size of the Moon but three times denser, may be pure iron and nickel but the outer core needs a small amount of lighter material to explain its density and low melting point. Mercury is also a dense planet, containing proportionately more iron than the Earth but Mars and the Moon contain substantially less iron than the Earth, even though they may have small cores. Based on its similar density and size Venus probably has an Earth-like core although the slightly lower pressures and possibly higher temperatures mean that a solid inner core may not be present.

The bulk of the Earth is contained in the mantle, the region between the core and the thin crust. The top of the mantle is sampled in several ways. Fragments are exposed in eroded mountain belts and brought to the surface by volcanic eruptions. The major mantle minerals excavated in these ways are olivine $(Mg,Fe)_2SiO_4$ and pyroxene $(Mg,Fe)SiO_3$, the main elements being magnesium, silicon and oxygen with iron being a minor constituent. These minerals are called silicates and they are also major constituents of meteorites. There is little calcium and aluminum in these minerals or in upper mantle rocks. These elements are abundant in the crust but the terrestrial crust is much too thin to give the Ca/Si or Al/Si ratios which are found in the Sun and the meteorites and, by inference, in planets. The most abundant material which emerges from the mantle is basalt. The floor of the ocean is covered with basaltic material and Hawaii and Iceland are two examples of thick basalt piles that have been built on the ocean floor. Hidden from view is the 80,000 km long chain of volcanoes defining the oceanic ridge system which generates new oceanic crust at the rate of 40 cubic kilometers per year. Basaltic magma is calcium and aluminum-rich and there must be vast quantities of basaltic material in the mantle. Basaltic magma is less dense than upper mantle material and erupts into or onto the crust. In fact, the majority of the crust was made in this way. However, cold basaltic material deeper than 60 km in the mantle converts to a very dense rock called eclogite. This is denser than shallow mantle rocks because it contains the dense aluminum-bearing crystal, garnet. Large bodies of eclogite can sink through the shallow mantle. This probably explains why the crust on Earth never gets thicker than about 60 km. On a smaller planet, the pressures are also smaller and the crust can be much thicker without converting to dense eclogite. On a hotter planet a thick basaltic crust would melt at its base, rather than convert to eclogite.

Although our direct samples of the Earth's interior are limited to the crust and shallow mantle we know from seismic tomography that there are broad low-velocity regions extending to a depth of at least 400 km under oceanic ridges and other volcanic terrains. Magmas and rocks containing magma have low densities and low seismic velocities and it is therefore reasonable to infer that the basalt source region is below about 400 km. The low density crystal mush ascends from great depth and then separates into melts which erupt at volcanoes and crystals which stay behind in the mantle or form new lithosphere.

Thus, we can identify three outer layers in the Earth; the buoyant crust, containing low density crustal minerals including quartz and feldspar; the shallow mantle containing high-temperature minerals and minerals that settle out of magma mushs rising to the surface - these minerals are primarily olivine and pyroxene; and a deep "fertile"-layer, containing a larger basaltic component and therefore abundant calcium and aluminum. This layer is dense when cold, because of the dense mineral garnet, and buoyant when hot, because garnet, and other easily melted minerals, melt to form basalt.

Seismic data suggests that the refractory (peridotite) layer extends to about 400 km depth and the basalt-eclogite-rich "fertile" layer occupies most of the remainder of the upper mantle and the shallow mantle under oceanic ridges. The lower mantle, which is the bulk of the mantle (70%), must be more silicon-rich than the upper mantle if the Earth has "cosmic" abundances of the elements. Primitive meteorites and the Sun have about one magnesium atom for every silicon atom. This favors the mineral enstatite $MgSiO_3$, a magnesium-rich pyroxene, over fosterite, Mg_2SiO_3, an olivine which contains two magnesium atoms for every silicon atom. The shallow mantle is olivine-rich having about two magnesiums per silicon. At high pressures the mineral olivine undergoes a series of phase changes and under lower mantle conditions it decomposes to a mixture of two minerals, MgO-periclase and $MgSiO_3$-perovskite. MgO has the rocksalt structure, the crystalline form of the alkali halides including ordinary rocksalt, NaCl. $MgSiO_3$-"perovskite" is the ultrahigh pressure form of enstatite and has the same crystal structure as the mineral perovskite ($CaTiO_3$) and many of the new high-temperature superconductors. "Perovskite" has much higher seismic velocities than periclase and matches the seismic velocities for the lower mantle. Therefore, it appears to be the most abundant mineral in the mantle. There is also some evidence from seismology that the lower mantle has more iron (as FeO) than the upper mantle, and may be similar to the mantles of the Moon and Mars, which are inferred to have FeO-rich mantles.

Where on Earth is the Crust?

The abundance of plagioclase, a light-weight and light-colored crustal silicate rich in calcium and aluminum, on the Moon is one evidence for a magma ocean. The average thickness of the lunar crust is greater than the Earth's crust even though the Moon is only 1% the size of the Earth. Does the absence of a thick crust on the Earth mean that it did not have a magma ocean, or that the crust mostly remained in or returned to the mantle? The lunar crust is so thick and contains such a large portion of the planets calcium and aluminum that it must have formed very efficiently, for example, by the floatation of light crustal minerals such as plagioclase to the top of a magma ocean. However, in a deep magma ocean on a larger body such as the Earth the pressures are so high that dense crystals such as garnet and pyroxene soak up the calcium and aluminum and they, by and large, stay in the mantle and may even sink to the base of a magma ocean. On the other hand some chemical elements, particularly the larger ones, are so concentrated in the Earth's crust that an efficient extraction mechanism must be postulated. These happen to be elements that are not particularly strongly partitioned into the high pressure minerals that replace plagioclase in a deep magma ocean. Planets grow by colliding with other objects, an energetic process that results in melting or even vaporization. Most of the energy is deposited in the outer layers, except for the small number of truly giant impacts which have as much chance of destroying a planet as of adding to it. These can melt a large fraction of a planet. A near surface magma ocean can process incoming material into light fractions, including melts, which float and dense materials, including refractory crystals and iron-rich melts which sink toward the center. The process can be compared to a blast furnace or a fat rendering plant. By heating and boiling, the original material is reduced to a frothy scum, the dense dregs and a "purified" liquid in between.

The inability of the mineral plagioclase to withstand very high pressure also explains why the Earth cannot have a thick "secondary" crust, that is, one formed by continental collision, mountain building or continuous addition of volcanic materials. When the Earth's crust becomes thicker than about 60 km not only do low density crustal minerals convert to denser minerals, causing the bottom of the crust to "fall off" (delaminate is the technical term) but even if it didn't the seismic velocities would be so high that a seismologist would call it mantle, not crust. "Crust", in fact, is a physical concept and its properties and thickness are derived from seismology. However, there are many samples of the lower crust and shallow mantle, provided by erosion and volcanism, and we know that the crust differs in chemistry, being calcium, aluminum and silicon-rich compared to the shallow mantle so that changes in physical properties are accompanied by changes in chemistry.

The crust on the Earth would be about 200 km thick if most of the low density and low melting point material had been separated from the dense and refractory material during Earth formation. Because of the high concentration of some elements in the crust we know that most of the mantle must have been processed or subjected to melt extraction either during accretion or shortly thereafter. It is therefore unlikely that the Earth was not very efficient in making crust but very likely that the missing crust resides somewhere in the mantle. The crust which is at the surface at any given time is only a small fraction of the total crustal material in the Earth and probably a small fraction of the total amount of crust that the Earth has made.

Physical State

Although most of the crust and mantle are solid we know from the seismic velocities, the abundance of volcanoes and from the increase of temperature with depth in wells and mines that much of the upper part of the Earth is near or above the melting point of rock. The coldest part of a planet is the surface and since cold rocks are strong and deform slowly we refer to the outer shell as the "lithosphere". At greater depth, averaging about 50 km, rocks become weak and behave as fluids, at least over geological time. The hot and weak portion of the upper mantle is called the "asthosphere". The asthenosphere is certainly hot and may be partially molten. Pressure operates in the opposite direction of temperature and rocks at greater depth in the Earth again become strong and harder to deform. Seismic velocities in the shallow mantle under young oceans and tectonic regions are so low that some partial melting is required to depths as great as 400 km.

The outer core is liquid, a result of high temperatures and the low melting point of iron alloys which melt at lower temperatures than common rocks. The viscosity of the Earth's core is very low, probably not much greater than water. Rapid motions of molten iron in the core are responsible for the Earth's magnetic field and for some of the jerkiness in the Earth's rotation. The density of the outer core is slightly less dense than pure molten iron and requires about 10% of some lighter elements such as S or O, or both. These elements are picked because they are abundant and they readily dissolve in the core.

The inner core of the Earth is solid, primarily the result of pressure-freezing (most liquids will freeze if the temperature is decreased or the pressure is increased). It "floats" in the center of the outer core and is essentially decoupled from the mantle. It represents only about 1.7% of the mass of the Earth.

The Lithosphere

Lithosphere implies "rocky" or "strong" layer and this is the name given to the outer shell of a planet. Rocks flow readily at the temperatures encountered in planetary interiors but, at the surface, the rocks are so cold and the viscosity so high that they support large loads and fail by brittle fracture rather than viscous flow. New lithosphere forms at midocean ridges and it thickens with time as it cools and moves away from the ridge. From the deflection pattern around large volcanoes and at trenches we know that the lithosphere acts as an elastic plate with thickness varying from zero at oceanic ridges to about 40 km for old ocean. As it cools, of course, it becomes denser and if it cools long enough it loses the high-temperature buoyancy it had at the ridge, and it will try to sink back into the interior.

The lithosphere is formed from the lighter components of mantle differentiation, melts and residual crystals. As the mantle convects, material in the hot upwellings partially melts and the denser crystals are left behind when the lighter material erupts. The lithosphere therefore may not be the same composition as normal mantle. The top layer, the oceanic crust, is basaltic and averages about 6 km in thickness. This is lighter than mantle material so this part of the lithosphere is buoyant. The top mantle layer of the lithosphere is composed of a rock called harzburgite which is essentially a normal mantle silicate minus the basaltic component. At moderate depth in the mantle, of the order of 50 km, basalt converts to the dense rock eclogite. This is denser than "normal" mantle and if it occurs in large enough blobs it will sink through the upper mantle. The melting of a "fertile peridotite" (harzburgite plus eclogite) eliminates the high density mineral garnet and both the melt and the refractory crystals are buoyant relative to unmelted or cold mantle and they will rise toward the surface to either form new lithosphere or to underplate previously formed lithosphere. We are ignorant of the composition of the lower oceanic lithosphere but it is probably a mixture of basalt, converting to eclogite as the lithosphere cools and thickens, and refractory crystals. Substantial amounts of dense eclogite in the older lithosphere would help explain why it eventually sinks back into the interior.

The continental lithosphere is about 150 km thick and both the crustal and mantle parts are buoyant relative to normal mantle. Continents therefore float around as icebergs and do not directly participate in the deeper circulatory part of mantle convection. In the course of continental drift continents can override the thinner oceanic lithosphere. Young and hot oceanic lithosphere tends to slide under the continent while thicker, colder, denser lithosphere tends to dive steeply into the mantle.

The lithosphere is an important element in planetary dynamics. If it gets too cold or too thick it can shut off the access of hot magmas to the surface and can be hard to break and subduct. If it contains too large a proportion of light minerals it will stay buoyant and will not sink back into the mantle. If there are too many plates or if they are moving rapidly then they again may not get dense enough to subduct. Thus, there are a variety of ways to choke up the surface of a planet. Plate tectonics can still exist on such a planet but any new crust and lithosphere that is formed at spreading centers must be compensated by lithospheric thickening or shallow underthrusting, the behavior of ice flows in the polar oceans. When the lithosphere gets too thick to break it can slide around as a unit on the underlying mantle, a one-plate planet. If a large volcano manages to form on the surface or if the surface is excavated by an impact, the moment of inertia of the planet can change and the whole outer shell will then rotate with respect to the spin axis.

There are several machanisms for breaking the lithosphere, a prerequisite for terrestrial style plate tectonics. Hot mantle upwellings can both heat and deform the lithosphere and divergent flow in the mantle places extensional stresses on the base of the lithosphere. Motion of the lithosphere over the surface of an elliptical rotating planet causes large stresses in the shell due to the changing shape of the surface. In particular, moving a plate from polar to equatorial regions causes extensional stresses. This motion of the surface relative to the spin axis can result from plate motions (apparent polar wander) or tilting of the whole planet (true polar wander). It is interesting that the supercontinent of Gondwanaland (Africa, South America, India, Australia, Antarctica) was at the South Pole 360 million years ago and broke up shortly after it had drifted across the equator (or after the tilting of the Earth had rotated it to the equator). Tidal despinning of a planet is a related method for generating large stresses in the surface layer and a global fracture pattern.

If the lithosphere becomes too dense it may sag and break. There are several subduction zones that are currently entirely in the ocean but it is not clear if they are the result of an instability of the lithosphere or if subduction started at the edge of a continent and migrated toward the ocean.

We do know that both oceanic ridges and island arcs can migrate relative to the underlying mantle and the spin axis.

Although plate tectonics, or at least "seafloor" spreading, may exist on Venus (see Venus chapter) that planet does not have the more obvious manifestations of terrestrial plate tectonics such as long linear ridges, subduction zones and deep trenches. Since Venus and Earth are so similar in size it is of interest to ask what differences might explain the difference in tectonic style. Venus is less rapidly rotating and therefore has a much smaller tidal bulge. The outer shell is nearly spherical and no large stresses are exerted as it moves around the surface. The surface temperature of Venus is much hotter than the Earth's surface. This makes for a thinner and more buoyant lithosphere, particularly if Venus has a crust similar to the Earth's. At the present time we do not know if it is the buoyancy of the lithosphere or the low spin rate and lack of a tidal bulge that make the tectonics of Venus different from the Earth. Evidence for some crustal extension, with short ridge segments and fracture zones, on Venus in equatorial latitudes has recently been found but there is, as yet, no evidence for subduction.

Plate Tectonics

When viewed from Africa the continents are drifting away from each other at rates of the order of 5-10 cm/year. About 180 million years ago the continents were assembled into a supercontinent called Pangea. The southern continents plus India were a single continental assemblage, Gondwana, for at least several hundred million years prior to that. About 360 million years ago this continent was centered on the South Pole but it moved toward the equator just prior to its breakup. Continental separation was initially rapid but slowed down as the separation increased. As the continents moved apart the

Atlantic Ocean opened up and the Pacific Ocean shrank. Part of the Pacific lithosphere disappeared underneath the continents.

Planets have various options for relieving themselves of their internal heat. The Earth chooses the plate tectonic option. Molten or semi-molten rock rises to the surface at midocean ridges, and their extensions into continents, where it cools rapidly and forms new oceanic crust. Most of the interior heat of the Earth is removed by this mechanism.

Plate tectonics begins by the creation of new crust and upper mantle at long linear globe encircling cracks called oceanic ridges. The associated volcanism is mostly underwater but the ridges can be traced around the world by their bathymetry and their seismic activity. As the newly formed lithosphere cools it contracts and moves away from the ridge, giving a characteristic smooth increase of ocean depth that is a function of age since formation, or distance from the ridge. This is called seafloor spreading. The ocean continues to deepen with age as the oceanic lithosphere cools further. The lithosphere eventually becomes unstable and sinks back into the mantle at subduction zones. These are linear or arcuate structures characterized by deep oceanic trenches and large volcanic cones capped by andesite.

By this process, the ocean floor continuously renews itself. Most of the ocean floor is less than 90 million years old and none is older than 200 million years. In 200 million years the oceanic lithosphere and shallow mantle has cooled to a depth of about 100 km and when this is inserted back into the hot mantle it acts as an icecube in a warm drink. This is the main mechanism by which mantle deeper than 100 km cools. Earthquakes occur to depths as great as 670 km and there is reason to believe that the cold oceanic plates, now called slabs, sink to this depth.

The Earth is apparently unique among the worlds we know about in having available to it this kind of cooling associated with deep subduction. Deep subduction can only occur if the temperature drop across the lithosphere is great enough to cause it to become unstable and to sink because of its greater density. On a planet with a thicker crust or a hotter surface or a colder interior the lithosphere may be permanently buoyant. In fact, on Earth the continents are permanently buoyant, a combination of thick low density crust and a buoyant upper mantle root extending to about 150 km depth.

The Earth actually exhibits at least three tectonic styles. The oceanic lithosphere recycles as just described. The continents are buoyant. They may break up and reassemble but they remain at the surface. However, they affect and are affected by the underlying mantle and adjacent plates. They are maintained against erosion - rejuvinated - by compressional mountain building and accretion of island arcs at their leading edge and by intrusion and extrusion of basalts, some of which may be trapped below and underplate the continent. Heat is removed from below a continent mostly by conduction, a relatively slow process. A continent can insulate the mantle and cause it to heat up and partially melt or, at least, protect the underlying mantle against the cooling action of subducted oceanic lithosphere. Since material flows from hot to cool parts of a convecting system continents will tend to drift away from hot mantle and come to rest over cool mantle. Most of the continents are now sitting on or moving toward cold parts of the mantle. The exception is the core of Pangea, Africa. The motion of continents may also force subduction of oceanic lithosphere. Many subduction zones are currently at the leading edges of continents and all of them could have formed at continental edges and migrated to their present positions.

A large supercontinent also prevents the cooling of the mantle below its interior by the subduction of cold oceanic lithosphere.

A small planet can cool more rapidly than a large planet and a given thickness lithosphere is harder to break up because of the lower gravity and less vigorous convection. Thus, the Moon (1% of mass of Earth) and probably Mars (10% of mass of Earth) are single plate planets and, with the exception of isolated volcanoes, must lose their interal heat by conduction. The interior is never exposed to the cooling effect of subducted lithosphere. The interior of such a planet can convect but the outer layer behaves as a more-or-less rigid shell. Hot upwellings in the interior can focus heat on one portion of the shell, weakening and thinning it, permitting magmas to rise to the surface and volcanoes to form, and even causing a limited amount of rifting and extension. On Earth, volcanic features called midplate volcanism, hotspots or plumes, account for about 10% of the heatflow from the interior. Thus, the Earth displays three different types of internal cooling.

There is another conceivable type of plate tectonics. If there is not a large temperature difference between the surface and the interior, of if plate generation is so rapid that deep cooling does not have a change to happen or if the crust-lithosphere system is buoyant then deep subduction cannot happen. If the plates are confined to the near surface then convergence zones result in deformation, plate thickening or shallow underthrusting and lithospheric doubling. This can be referred to as "pack-ice" tectonics. Venus, and the early Earth may experience this tectonic style. Divergence zones of crustal generation occur but convergence zones are diffuse and characterized by high elevation. Western North America, parts of western South America and Tibet may be terrestrial examples of this buoyant plate or pack-ice tectonics.

Geoid

On an entirely fluid planet the shape of the surface is controlled by rotation and the gravitational attraction of mass anomalies in the interior. Regions of mass excess attract the fluid and the surface stands high. This shape is called the geoid. The surface of the ocean approximates the geoid. Satellite techniques are used to calculate the geoid for the whole Earth (Figure). The geoid is usually expressed in terms of spherical harmonics and referenced to some shape such as an ellipsoid of revolution so that the rotational deformation is removed. The resulting figure has undulations of the order of hundreds of meters. Although the distribution of density anomalies in the interior cannot be unambiguously determined from gravity, or geoid, data alone it is possible to calculate the contribution from isostatically compensated continents, slabs and lower mantle density variations.

At very long wavelength there are equatorial geoid highs centered on the Pacific and on Africa. Geoid lows occur in a polar band extending through North America, Brazil, Antarctica, Austria and Asia. Hager, Clayton and Dziewonski have shown that this pattern correlates with the seismic velocity distribution in the lower mantle. Hot mantle is expected to have low seismic velocities and low densities and these are expected to occur in the hot upwellings in a convecting mantle. At long wavelengths these upwellings cause geoid highs because of the upward deformation of the core-mantle boundary and the surface.

Except for Africa the continents are in or near geoid lows. They give the impression of having migrated into these regions as they moved away from Africa after the destruction of Pangea.

At shorter wavelengths subduction zones show up as geoid highs, or mass excesses. This is expected if slabs are cold and dense, as long as they are supported from below by a strong, or dense or highly viscous lower mantle.

There is poor correlation of present day tectonics or elevation with the geoid. However, there is good correlation with continental and subduction zone configurations in the past. For example, the geoid high centered over Africa has about the shape and size of Pangea and geoid lows correspond roughly with inferred regions of subduction prior to extensive opening of the Atlantic Ocean. This may be a result of continental insulation by the supercontinent, and adjacent cooling by subduction.

The major geoid highs of moderate wavelength are associated with subduction in New Guinea to Tonga and Peru-Chile and these are centered on the equator, undoubtedly contributing to the moment of inertia that controls the orientation of the spin axis relative to the mantle.

Table 1. Summary of Earth Structure

Region	Depth (km)	Fraction of Total Earth Mass	Fraction of Mantle and Crust
Continental crust	0-50	0.00374	0.00554
Oceanic crust	0-10	0.00099	0.00147
Upper mantle	10-400	0.103	0.153
Transition region	400-650	0.075	0.111
Lower mantle	650-2890	0.492	0.729
Outer core	2890-5150	0.308	--
Inner core	5150-6370	0.017	--

Table 2

The amount of various elements in the crust relative to the total estimated amount in the Planet (%)

Rb	68	Sr	21
Cs	67	Na	13
Th	55	Al	2.4
Ba	49	Ca	0.9
U	47	Si	0.7
La	27	Mg	0.1

Table 3

Estimated compositions of planets which contain the major planetary elements (Si, Mg, Fe, Al and Ca) in solar or chondritic proportions plus the oxygen required to oxidize these elements. Enough Fe is removed from the mantle to form an Earth sized core (32.5 weight percent of the planet) when combined with enough oxygen (Fe_2O) to give the observed density. Components are given in weight percent.

	Chondritic	Solar
MANTLE		
MgO	37.7	32.7
SiO_2	52.5	45.0
Al_2O_3	3.8	3.2
CaO	3.0	3.4
FeO	3.3	15.7
CORE		
Fe_2O	32.2	32.0

Suggested Figures

1. Earth topography (Washington University)

2. Seasat Geoid (Haxby, JPL or Rapp)

3. Long wavelength geoid (Hager)

4. Seismic tomography maps, several depths

5. Seismic tomography cross-sections

6. Space photos or images of selected features

    1. Hotspots (Hawaii, Iceland)

    2. Subduction zone volcanoes (radar image of Shasta (JPL), Japan, Aleutions)

    3. Transform fault; San Andreas, Alpine, N.Z.

    4. Triple junction, hot spot - Afar

7. Tomography cross-section of slab

8. Cartoon cross section of upper mantle

9. Plate tectonic map

10. Cut away cross section of entire globe (Harvard)

11. Plus Earth figures from previous editions.

Figure Captions

Figure 1 Seismic shear velocity in the mantle at a depth of 250 km as determined from surface wave tomography. The red regions have low seismic velocities and are therefore hot. The blue regions are colder. Note the association of hot mantle with oceanic ridges and continental tectonic regions (from a Caltech model of Nataf, Nakanishi and Anderson).

Figure 2 Shear velocity at a depth of 380 km. Note the appearance of a hot anomaly in the central Pacific. The white circles are hotspots and the white lines are plate boundaries. (Caltech model as above)

Figure 3 Compressional velocity at a depth of 1000 km. The red regions are presumably hot upwellings (from a Harvard model of Dziewonski and Woodhouse).

Figure 4 Same as above at a depth of 2500 km (Harvard model as above)

Figure 5 Seismic tomographic cross-section across the subduction zone of N. Japan. The red regions have low compressional velocities and are therefore hot. The stars are earthquakes and they fall near the top of a cold (blue) slab. (Caltech model of Zhou and Clayton)

Figure 6 Seismic cross section across the path shown in the upper figure, from a depth of 50 to 550 km. The open circles are slow, the bigger the slower. The size of the closed circles is proportional to the velocity of the shear waves, in excess of the average velocity. Note the low velocities, and therefore hot temperatures under Saudi Arabia and the younger parts of the oceans. Dashed lines are plate boundaries. The cross sections are shown at two different vertical exaggerations.

Figure 7 Cartoon of the plate tectonic cycle.