Important Concepts in
Geophysics, Geology and Volcanology

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First Edition, 2012

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Mantle Plume
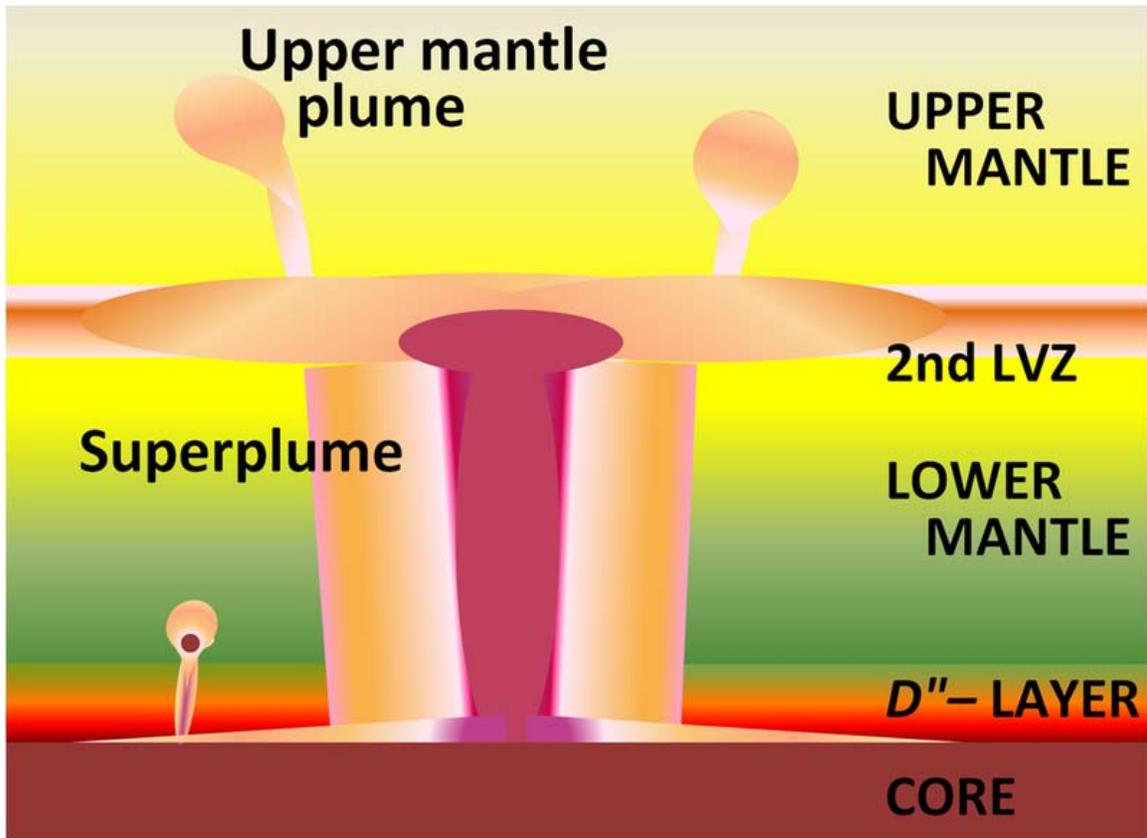
A superplume generated by cooling processes in the mantle

A **mantle plume** is an upwelling of abnormally hot rock within the Earth's mantle. As the heads of mantle plumes can partly melt when they reach shallow depths, they are often invoked as the cause of volcanic centers known as hotspots and their related flood basalts, though there is some scientific debate about this. They may be far from tectonic plate boundaries.

A mantle plume is a secondary way that Earth loses heat, much less important in this regard than is heat loss at plate margins. Some scientists think that plate tectonics cools the mantle, and mantle plumes cool the core.

The geometry of the Hawaiian-Emperor seamount chain and the regular progression of ages of volcanism along it were taken as important evidence in support of the mantle plume theory.

Concepts

In 1971, geophysicist W. Jason Morgan proposed the theory of mantle plumes. In this theory, convection in the mantle slowly transports heat from the core to the Earth's surface. It is now understood that two convective processes drive heat exchange within the earth: *plate tectonics*, which is driven primarily by the sinking of cold plates of lithosphere back into the mantle asthenosphere, and *mantle plumes*, which carry heat upward in rising columns of hot material, driven by heat exchange across the core-mantle boundary. The sinking of vast sheets of oceanic lithosphere back into the mantle is the primary driving force of plate tectonics, where the sinking of these slabs is balanced by the passive upwelling of asthenosphere along mid-oceanic ridges. In contrast, mantle plumes are narrow columns of material that rise more-or-less independently of plate motions.

Fluid dynamics experiments in the early 1970s produced models of mantle plumes that consist of two parts: a long thin conduit that connects the top of the plume to its base, and a bulbous head that expands in size as the plume rises upward in the mantle; the result looks like a mushroom with a thin stalk and large top. The bulbous head forms because hot material moves upward through the plume conduit faster than the plume itself rises through the surrounding asthenosphere. In the late 1980s and early 1990s, experiments with thermal models shows that as the bulbous head expands it may entrain some of the adjacent asthenosphere into the rising head.

When the plume head encounters the base of the lithosphere, it flattens out against this surface barrier and undergoes widespread decompression melting to form enormous volumes of basalt magma. This basalt may erupt onto the surface over very short time scales (less than 1 million years) to form a continental flood basalt (if it erupts through continental crust) or an oceanic plateau (if it erupts through oceanic crust). Prominent continental flood basalt provinces include the Deccan traps and the Rajmahal traps in India, the Siberian traps of Asia, the Karmutsen Formation in British Columbia, Canada, the Karoo-Ferrar basalts/dolerites in South Africa and Antarctica, the Paraná and Etendeka traps in South America and Africa (formerly a single province separated by opening of the South Atlantic Ocean), and the Columbia River basalts of North America. Plume-related oceanic plateaux include the Ontong Java plateau of the southwest Pacific Ocean and the Maniheken plateau of the Indian Ocean.

The plume tail may continue to move material from the Earth's interior to the surface, providing a continuous supply of magma in a fixed location, often referred to as a hotspot. As the lithosphere moves over this fixed hotspot due to plate tectonics, the eruption of magma from the fixed hotspot onto the surface forms a chain of volcanoes that parallels plate motion. The classic example of this is the Hawaiian Islands chain in the Pacific Ocean.

The eruption of continental flood basalts is often associated with continental rifting and breakup, leading to the hypothesis that mantle plumes play an important role in continental rifting and the formation of ocean basins. Where this association of flood basalts with continental rifting is observed, it is not uncommon to find linear chains of volcanic islands that parallel the motion of plates on either side of the spreading center (South Atlantic Ocean).

Model of plume formation
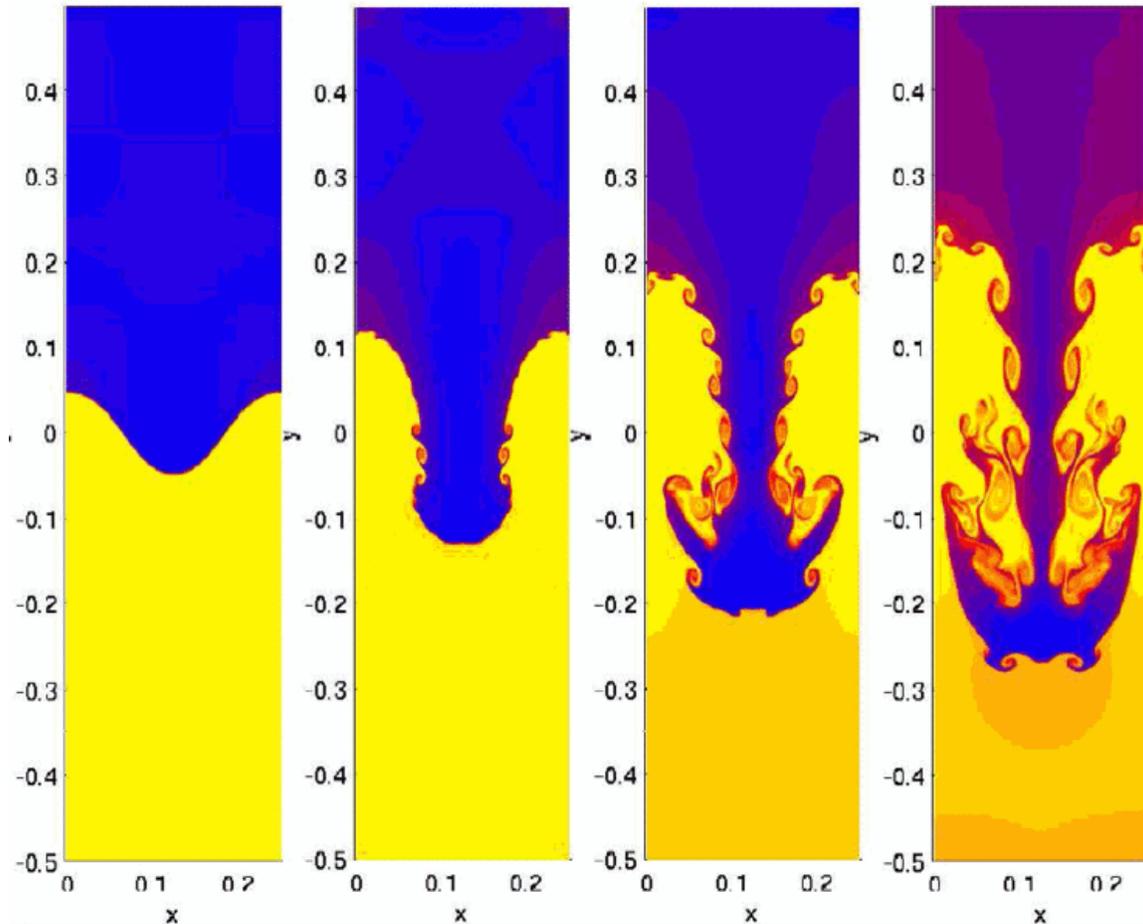
Hydrodynamic simulation of a single "finger" of the Rayleigh–Taylor instability, a possible mechanism for plume formation. Note the formation of a "mushroom cap" at a later stage in the third and fourth frame in the sequence.
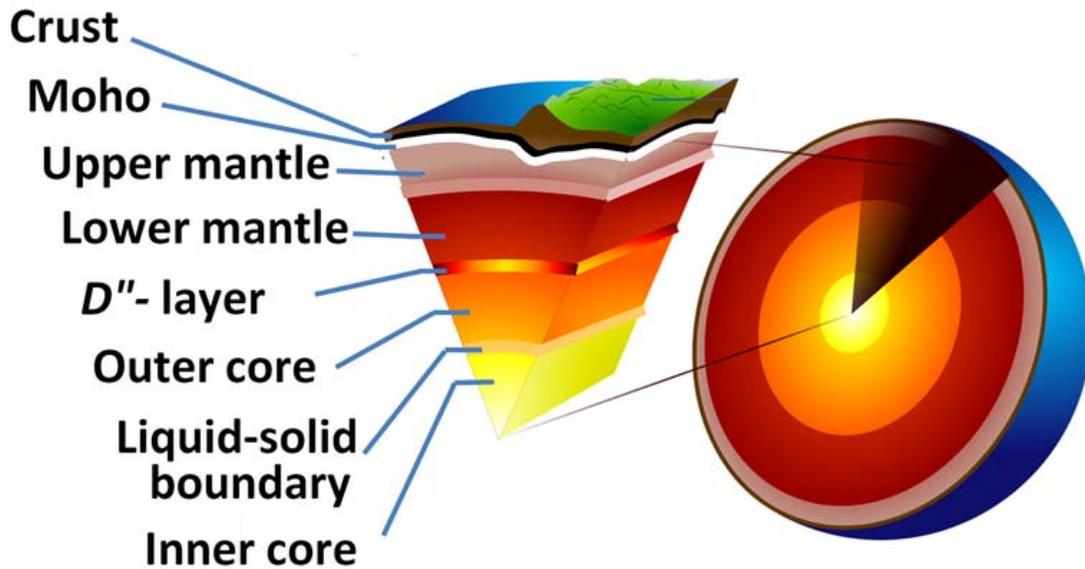
Earth cross-section showing location of upper and lower mantle, D''-layer, and outer and inner core

The chemical and isotopic composition of basalts found in hotspots and inferred to form by partial melting of mantle plumes suggest that several components are involved, including primordial mantle with unfractionated noble gases, subducted oceanic crust and mantle lithosphere, and subducted sediments. The processing of oceanic crust, lithosphere, and sediment through a subduction zone decouples the water soluble trace elements (e.g., K, Rb, Th) from the immobile trace elements (e.g., Ti, Nb, Ta), concentrating the immobile elements in the oceanic slab (the water soluble elements are added to the crust in island arc volcanoes). Seismic tomography shows that subducted oceanic slabs may sink directly to the core-mantle boundary, or pause for long periods at the mantle transition zone (400–660 km depth) before sinking to the core-mantle boundary. The subducted slabs accumulate at the core-mantle boundary and form a seismically distinct layer called the D'' (Dee-double prime). This appears to be the source of most deep mantle plumes, as shown by seismic tomography.

Because there is little material transport across the core-mantle boundary, heat transfer must occur by conduction, with well-stirred adiabatic gradients above and below this boundary. As a result, the core-mantle boundary represents a significant thermal (temperature) discontinuity, with the core at temperatures several hundred degrees Celsius hotter than the overlying mantle. As heat is transferred across this boundary by conduction, material in the D" layer becomes hotter and thus more buoyant. When it becomes sufficiently buoyant, material begins to rise from the D" layer to form a mantle plume.

In concert with hypothesised slow-down in plate tectonic motion, which may be associated with prolonged periods of supercontinent formation, it is theorised that without an actively convecting asthenosphere, the lower mantle will begin to locally overheat.

These overheated portions of the mantle near the core-mantle boundary become buoyant relative to their surroundings, and begin to rise via diapirism.

This plume of material rises through the mantle. Upon reaching shallower depths within the asthenosphere, decompression melting occurs in the plume head, creating large volumes of magma. The magma rises through the asthenosphere until it reaches the Earth's crust where it causes a hotspot.

Miniplumes

The term *mini-plumes* refers to smaller plumes that may originate in the upper mantle rather than the more common deep mantle plumes. No conclusive examples have yet been identified. One possible example, however, is the Anahim plume at the Anahim hotspot in central British Columbia, Canada.

Role of the core
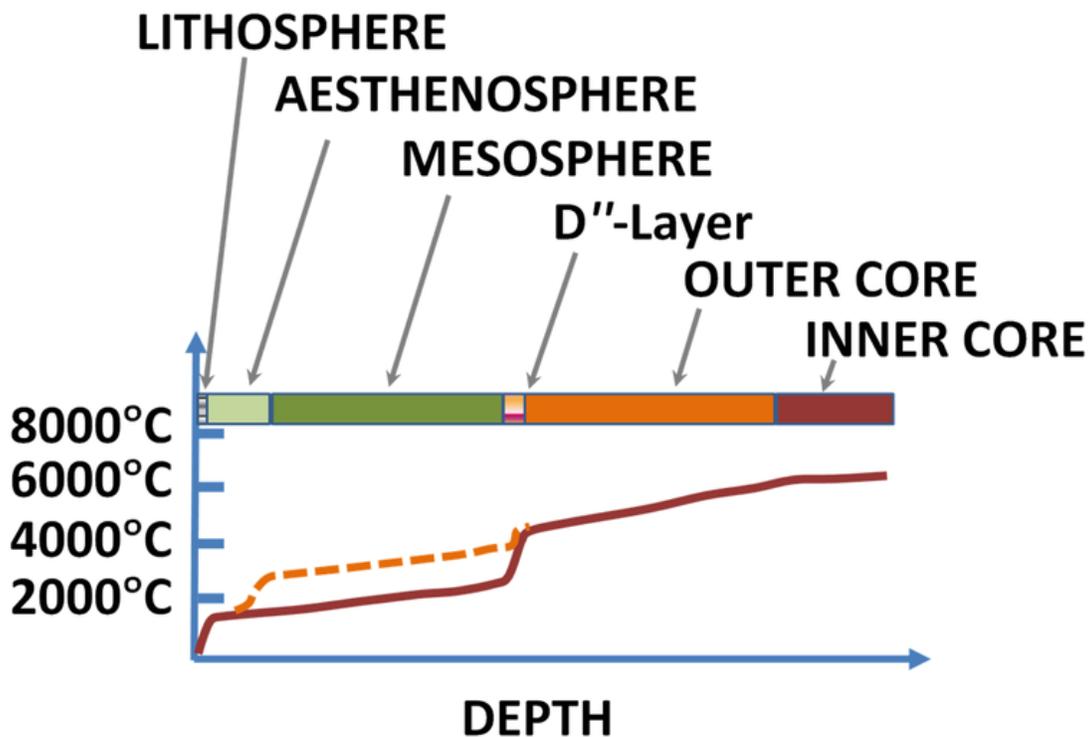
Calculated Earth's temperature vs. depth. Dashed curve: Layered mantle convection; Solid curve: Whole mantle convection.

The most prominent compositional contrast known to exist in the deep ($\gtrsim$ 400 km) mantle is at the core-mantle boundary. *Morgan-type plumes* are generally assumed to rise from this layer for two reasons. First, this boundary represents a major thermal

discontinuity because the top of the core is much hotter than the base of the mantle. Secondly, the base of the mantle is characterized by the D″-layer that is seismically distinct from the overlying mantle. The D″-layer appears to be compositionally distinct from the overlying mantle, and seismic tomography of subducted lithosphere suggests that the D″ layer may represent the accumulation of these subducted slabs at the base of the mantle.

Very large, broad plumes that spawn a series of smaller plumes in the upper mantle are sometimes referred to as "superplumes". These are usually defined as a plume that has a diameter of at least 1500–3000 km by the time the plume head reaches the upper mantle. A "superplume event" is a short-lived mantle event (100 million years) during which a superplume and the smaller plumes that form from it bombard the base of the lithosphere. It is believed that such an event may have occurred in the mid-Cretaceous.

Evidence for the theory

Mantle plumes provide an explanation for intra-plate tectonic volcanism called 'hotspots'. There are several lines of evidence used to support the theory: linear volcanic centers, hotspot fixity, geochemical, noble gas isotopes, and geophysical anomalies.

Linear volcanic tracks

The apparent linear, age-progressive distribution of the Hawaiian-Emperor seamount chain is explained in this context as a result of a fixed, deep-mantle plume impinging into the upper mantle, partly melting, and causing a "track" as the plate moves with respect to the plume source.

Smaller plumes, arguably called petitspots, are also common within intraplate areas. For instance, tracks of ocean island basalts are found within the Indian Plate, namely the Marshall Islands hotspot.

Continental flood basalt in Oregon and Washington and the Yellowstone caldera-forming event are also used as evidence for mantle plumes, with the voluminous flood basalt envisaged as a product of the vigorous mantle plume head, and the hot 'tail' to the plume driving a progressively younger series of caldera events as the North American continental mass tracks above it.

Smaller series of intracontinental volcanic rocks are also ascribed to small plumes or petitspots. These are notably the Glasshouse Mountains in Queensland (Cohen et al. 2004), which are the oldest Tertiary (25 Ma) members of a progressively younger trend of basaltic and intraplate volcanic cones and plugs culminating in the maars and small peridotitic basalts of the Newer Volcanics in Victoria of 40,000 years ago, far to the southeast.

It is notable that these volcanic features become younger in the same vector as the motion of the Indo-Australian Plate, and matching the trend of the intraplate ocean island basalts in the Indian Ocean.

Noble gas and other isotopes

The standard ^{3}He is considered a primordial isotope as it was formed in the Big Bang and very little is produced or added to the Earth by other processes since then (Anderson, 1989). ^{4}He includes a primordial component, but it is also produced by the natural radioactive decay of U and Th. ^{3}He is 25% lighter than ^{4}He, so over time He in the upper atmosphere becomes depleted in ^{3}He as it is lost into space. All of these processes contribute to low ratios of ^{3}He/^{4}He in the atmosphere and in the Earth's crust and upper mantle. Thus, the only potential source of He with elevated ^{3}He/^{4}He ratios is the deep interior of the Earth, which must still remain a reservoir of primordial He and other gases. Thus, anomalous ^{3}He/^{4}He isotopic ratios with respect to mean mid-ocean ridge basalts (MORB), as found in Hawaiian volcanic rocks, are assumed to provide a signature of primordial, non-degassed mantle (however: alternate explanations have been proposed for this anomalous geochemical signature.

Relative abundances of osmium isotopes in Hawaiian basalts have also been taken as signatures of plume formation at the core-mantle boundary, with incorporation of some core-derived material. That explanation for the osmium isotope abundances remains controversial.

Geophysical anomalies
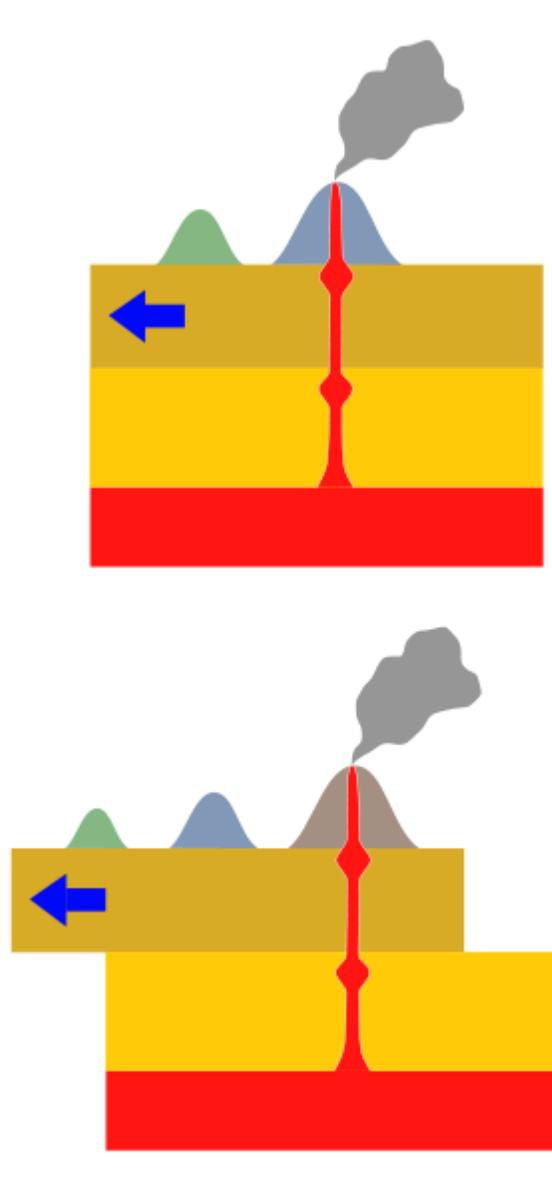
Diagram showing a cross section though the Earth's lithosphere (in yellow) with magma rising from the mantle (in red). The crust may translate relative to the plume, creating a *track*.

Geophysical anomalies associated with hotspots and plumes include thermal, seismic, and geodetic. Thermal anomalies are inherent in the term "hotspot." Thermal anomalies are reflected in high heat flow values at the Earth's surface and excess volcanism. Thermal anomalies also produce anomalies in the travel times of seismic waves.

Seismic anomalies are identified by measuring spatial variations in the time it takes seismic waves to travel through the earth. A fluid body with a lower density (e.g., a hot

mantle plume or wetter mantle) exhibits lower seismic velocity compared to surrounding mantle. Observations of regions where seismic waves take longer to arrive are used as evidence for regions of anomalously hot mantle, as is observed underneath Hawaii. Other indicators of plumes would be from the dynamic uplift of the surface and an elevated heat flow.

By deploying a dense network of seismometers and a technique known as seismic tomography, scientists can construct 3-d images of seismic velocities to try and identify vertical plume like structures. This is referred to as *seismic tomography* because it uses techniques similar to medical tomography.

Seismic waves generated by large earthquakes are used to determine structure below the Earth's surface because they can be detected far from the earthquake epicenter. Far-travelled seismic waves (also called teleseismic waves) are especially useful for seismic tomography because they have steep travel paths that sample smaller longitudinal domains. Density differences between a mantle plume and cooler material that surrounds it enable researchers to distinguish between the two. Seismic waves slow down when they travel through low-density (hotter) material, and speed up when traveling through denser (cooler) material. Density differences may also arise from compositional differences between the plume material and the surrounding mantle.

By analyzing pressure pulses, or P-waves, a group of scientists at Princeton have identified 32 regions throughout the world where P-waves travel slower than average. They conclude that these areas are mantle plumes. The team used analysis of S-waves, another type of seismic wave generated by earthquakes, to determine that those plumes extend to the core-mantle boundary.

Geodetic anomalies are reflected in topographic bulges above the plume location, and in positive geoid anomalies. The geoid is a potential surface that reflects the theoretical height to sealevel if mass was distributed uniformly within the Earth. Positive geoid anomalies reflect excess mass associated with uplift and doming over a thermal plume. The Yellowstone plume has a positive geoid anomaly of around +15 meters at its center, and over 1000 km in diameter.

Computer modeling of the mantle plume theory shows that changes of temperature and chemical composition of rising plumes can lead to plumes of varying contours as opposed to the early conceptualization that plumes developed as a homogeneous mushroom shape.

Geochemistry

Basalts associated with hotspots or mantle plumes are geochemically distinct from mid-ocean ridge basalts and from lavas associated with island arc volcanoes. In major elements, hotspot basalts are typically higher in iron (Fe) and titanium (Ti) than mid-ocean ridge basalts at similar magnesium (Mg) contents, reflecting their higher temperatures of formation. In trace elements, hotspot basalts are typically more enriched

in the light rare earth elements than mid-ocean ridge basalts. Compared to island arc basalts, hotspot basalts are lower in alumina (Al2O3) and much higher in the immobile trace elements (e.g., Ti, Nb, Ta).

The significance of these differences among ocean island basalts (hotspots), mid-ocean ridge basalts, and island arc basalts rests on processes that occur during subduction of oceanic crust and mantle lithosphere. Oceanic crust (and to a lesser extent, the underlying mantle) typically becomes hydrated to varying degrees on the seafloor, partly as the result of seafloor weathering, and partly in response to hydrothermal circulation near the ridge crest. As oceanic crust-lithosphere subduct, water is released by dehydration reactions, along with water-soluble chemical elements and trace elements. This enriched fluid rises to metasomatize the overlying mantle wedge and leads to the formation of island arc basalts. The subducting slab is depleted in these water-mobile elements (e.g., K, Rb, Th, Pb) and thus relatively enriched in elements that are not water-mobile (e.g., Ti, Nb, Ta) compared to both mid-ocean ridge and island arc basalts.

Ocean island basalts, which represent the volcanic product of mantle plumes, are also relatively enriched in the immobile elements relative to the water-mobile elements, leading to the conclusion that subducted oceanic crust plays a major role in their origin.

Suggested mantle plume locations

Two of the most well known locations that fit the mantle plume theory are Hawaii and Iceland as both have volcanic activity. Other island chains that parallel plate motion include the Society Islands (e.g., Tahiti), St. Helena-Ascension-Gough, and the Ninety-east ridge (Indian ocean). One of the dormant in Asia that fits the mantle plume theory is Mount Halla (Hallasan) in Jeju island (Jeju-do).

The P-wave and S-wave images show other locations that fit the mantle plume model. Ascension Island and St. Helena appear to originate from the same plume. Similarly, volcanic activity in the Azores and Canary Islands branch from a single trunk.

South of Java and in the Coral Sea, the images show possible formation of future plumes that currently extend only halfway to the surface.

The Mississippi Embayment in North America which formed in the mid-Cretaceous has been suggested as a mantle plume event that occurred as the North American plate, specifically the Mississippi graben, passed over the Bermuda hotspot.

Ore deposit associations

- Nickel-Copper-PGE deposits. For instance the giant Norilsk nickel deposit in Russia is considered to be associated with the Permian Siberian Traps volcanism, a probable plume-head eruptive event.
- Gold deposits (to a lesser extent)

Alternative models of hotspot formation

It is important to distinguish between observation and interpretation or hypothesis. Hotspots are observed surface features characterized by volcanic effusions in excess of what is normally expected for their nominal setting; mantle plumes are interpreted to be the cause of many or most hotspots. In a 2005 paper, Don L. Anderson and James H. Natland wrote:

> "Unfortunately, the terms hotspot and plume have become confused. In recent literature the terms are used interchangeably. A plume is a hypothetical mantle feature. A hotspot is a region of magmatism or elevation that has been deemed to be anomalous in some respect because of its volume or location. In the plume hypothesis, a hotspot is the surface manifestation of a plume, but the concepts are different; one is the presumed effect, and the other is the cause."

Although mantle plumes are currently the dominant hypothesis for creating hotspots, flood basalts, and oceanic plateaux, collectively referred to as large igneous provinces (LIPS), many geoscientists prefer models that are confined to the upper mantle and crust, and do not require deep thermal anomalies. These hypotheses include:

- Crustal delamination: the delamination and sinking of large portions of lower continental crust (assumed to have transformed into the dense rock eclogite), which allows the influx of asthenosphere from the low velocity zone and subsequently melts to form continental flood basalts.

- Edge effects: thick continental lithosphere insulates the underlying asthenosphere, causing heat buildup that leads to buoyancy. The buoyant asthenosphere moves toward the edge of the cratonic lithosphere, where it can rise and melt (e.g., Anderson, 2005). This model is often paired with rifting of the continental crust to explain formation of ocean basins.

- Meteorite impacts: the impact of large meteorites into oceanic crust may cause large parts of the transient cavity to melt, forming melt sheets similar in volume to flood basalts. This model works less well in continental crust, which lacks a basaltic bulk composition.

The current debate has stimulated an increased interest in research to distinguish between these models. Recent advances in seismic tomography have enhanced its spatial resolution in both the upper and lower mantle. New seismic tomographs resolve anomalous features consistently within the upper mantle, and in places to the lower mantle. It is becoming difficult to explain these data by processes in the uppermost mantle.

Chapter- 2

Eruption Column

Eruption column over Mount Pinatubo in the Philippines

An **eruption column** consists of hot volcanic ash emitted during an explosive volcanic eruption. The ash forms a column rising many kilometres into the air above the peak of the volcano. In the most explosive eruptions, the eruption column may rise over 40 km, penetrating the stratosphere. Stratospheric injection of aerosols by volcanoes is a major cause of short-term climate change.

A common occurrence in explosive eruptions is for *column collapse* to occur. In this case, the eruption column is too dense to be lifted high into the air by air convection, and instead falls down the flanks of the volcano to form a pyroclastic flow or surge.

Formation

Eruption columns form in explosive volcanic activity, when the high concentration of volatile materials in the rising magma causes it to be disrupted into fine volcanic ash and coarser tephra. The ash and tephra are ejected at speeds of several hundred metres per second, and can rise rapidly to heights of several kilometres, lifted by enormous convection currents.

Eruption columns may be transient, if formed by a discrete explosion, or sustained, if produced by a continuous eruption or closely spaced discrete explosions.

Structure

The solid or liquid material in an eruption column is lifted by processes which vary as the material ascends:

- At the base of the plume, material is forced upwards out of the vent by the pressure of expanding gas, mainly steam. The gas expands because the pressure of rock above it rapidly reduces as it approaches the surface. This region is called the *gas thrust region* and typically reaches to only one or two kilometres above the vent.

- The *convective thrust region* covers most of the height of the plume. The gas thrust region is very turbulent and surrounding air becomes mixed into it and heated. The air expands, reducing its density and rising. The rising air carries the solid and liquid material from the eruption entrained in it upwards.

- As the plume rises into less dense surrounding air, it will eventually reach an altitude where the hot, rising air is of the same density as the surrounding cooler air. In this neutral buoyancy region, the erupted material will then no longer rise through convection, but solely through any upward momentum which it has. This is called the *umbrella region*, and is usually marked by the column spreading out sideways. The eruptive material and the surrounding cool air has the same density at the base of the umbrella region, and the top is marked by the maximum height which momentum carries the material upward. Because the speeds are very low or negligible in this region it is often distorted by stratospheric winds.

Column heights

Eruption column rising over Redoubt volcano, Alaska

The column will stop rising once it reaches an altitude where it is no longer less dense than the surrounding air. Several factors control the height that an eruption column can reach.

Intrinsic factors include the diameter of the erupting vent, the gas content of the magma, and the velocity at which it is ejected. Extrinsic factors can be important, with winds sometimes limiting the height of the column, and the local thermal temperature gradient also playing a role. The atmospheric temperature in the troposphere normally decreases by about 6-7 K/km, but small changes in this gradient can have a large effect on the final column height. Theoretically, the maximum achievable column height is thought to be about 55 km. In practice, column heights ranging from about 2-45 km are seen.

Eruption columns over 10-15 km high break through the tropopause and inject ash and aerosols into the stratosphere. Ash and aerosols in the troposphere are quickly removed by rain and other precipitation, but material injected into the stratosphere is much more slowly dispersed, in the absence of weather systems. Substantial amounts of stratospheric injection can have global effects: after Mount Pinatubo erupted in 1991, global temperatures dropped by about 0.5°C. The largest eruptions are thought to cause drops of up to several degrees, and are potentially the cause of some of the known mass extinctions.

Eruption column heights are a useful way of measuring eruption intensity since for a given atmospheric temperature, the column height is proportional to the fourth root of the mass eruption rate. Consequently, given similar conditions, to double the column height requires an eruption ejecting 16 times as much material per second. The column height of eruptions which have not been observed can be estimated by mapping the *maximum* distance that pyroclasts of different sizes are carried from the vent—the higher the column the further ejected material of a particular mass (and therefore size) can be carried.

The approximate maximum height of an eruption column is given by the equation.

$$H = k(M\Delta T)^{1/4}$$

Where:

k is a constant that depends on various properties, such as atmospheric conditions.
M is the mass eruption rate.
ΔT is the difference in temperature between the erupting magma and the surrounding atmosphere.

Hazards

Column collapse

Eruption columns may become so laden with dense material that they are too heavy to be supported by convection currents. This can suddenly happen if, for example, the rate at which magma is erupted increases to a point where insufficient air is entrained to support it, or if the magma density suddenly increases as denser magma from lower down in a stratified magma chamber is tapped.

If it does happen, then material reaching the bottom of the convective thrust region can no longer be adequately supported by convection and will fall under gravity, forming a pyroclastic flow or surge which can travel down the flanks of a volcano at speeds of over 100 km/h. Column collapse is one of the most common and dangerous volcanic hazards in a plinian eruption.

Aircraft

Several eruptions have seriously endangered aircraft which have encountered the eruption column. In two separate incidents in 1982, an airliner flew into the upper reaches of an eruption column generated by Mount Galunggung, and the ash severely damaged both aircraft. Particular hazards were the ingestion of ash stopping the engines, the sandblasting of the cockpit windows rendering them largely opaque and the contamination of fuel through the ingestion of ash through pressurisation ducts. The damage to engines is a particular problem since temperatures inside a gas turbine are

sufficiently high that volcanic ash is melted in the combustion chamber, and forms a glass coating on components further downstream of it, for example on turbine blades.

In one case, the aircraft lost power on all four engines, and in the other, three of the four engines failed. In both cases, engines were successfully restarted but the aircraft were forced make emergency landings in Jakarta.

Similar damage to aircraft occurred due to an eruption column over Redoubt volcano in Alaska in 1989. Following the eruption of Mount Pinatubo in 1991, aircraft were diverted to avoid the eruption column, but nonetheless, ash dispersing over a wide area caused damage to 16 aircraft, some as far as 1000 km from the volcano.

Eruption plumes are not usually visible on weather radar and may be obscured by cloud or night. Because of the risks posed to aviation by eruption plumes, there is a network of nine Volcanic Ash Advisory Centers around the World which continuously monitor for eruption plumes using data from satellites, ground reports, pilot reports and meteorological models.

Chapter- 3

Magma

Lava flow on Hawaii. Lava is the extrusive equivalent of magma

Magma (from Greek μάγμα "paste") is a mixture of molten rock, volatiles and solids that is found beneath the surface of the Earth, and may also exist on other terrestrial planets. Besides molten rock, magma may also contain suspended crystals and gas bubbles. Magma often collects in magma chambers that may feed a volcano or turn into a pluton. Magma is capable of intrusion into adjacent rocks, extrusion onto the surface as lava, and explosive ejection as tephra to form pyroclastic rock.

Magma is a complex high-temperature fluid substance. Temperatures of most magmas are in the range 700 °C to 1300 °C (or 1300 °F to 2400 °F), but very rare carbonatite melts may be as cool as 600 °C, and komatiite melts may have been as hot as 1600 °C. Most are silicate mixtures.

Environments of magma formation and compositions are commonly correlated. Environments include subduction zones, continental rift zones, mid-oceanic ridges, and hotspots, some of which are interpreted as mantle plumes. Despite being found in such widespread locales, the bulk of the Earth's crust and mantle is not molten. Rather, most of the Earth takes the form of a rheid, a form of solid that can move or deform under pressure. Magma, as liquid, preferentially forms in high temperature, low pressure environments within several kilometers of the Earth's surface.

Magma compositions may evolve after formation by fractional crystallization, contamination, and magma mixing. By definition, all igneous rock is formed from magma.

While the study of magma has historically relied on observing magma in the form of lava outflows, magma has been encountered in situ three times during drilling projects—twice in Iceland, and once in Hawaii.

Source

Partial melting
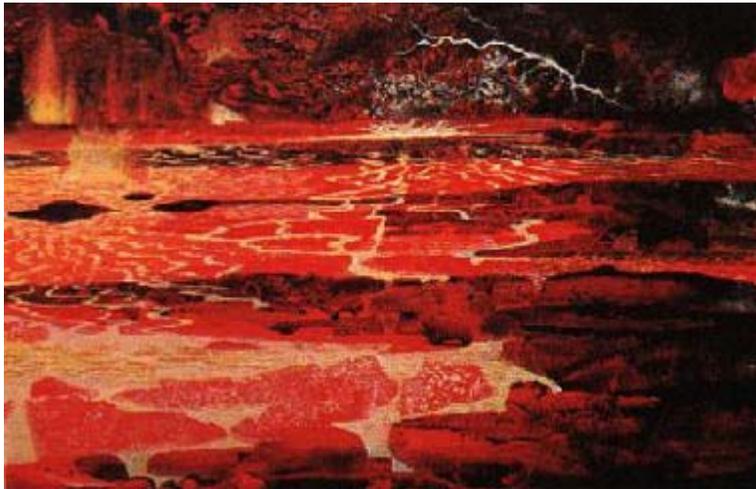
An artist's impression of a magma ocean on the early Earth

Melting of solid rock to form magma is controlled by three physical parameters: its temperature, pressure, and composition. Mechanisms are discussed in the entry for igneous rock.

When rocks melt they do so incrementally and gradually; most rocks are made of several minerals, all of which have different melting points, and the phase diagrams that control melting are often complex. As a rock melts, its volume changes. When enough rock is melted, the small globules of melt (generally occurring in between mineral grains) link up

and soften the rock. Under pressure within the earth, as little as a fraction of a percent partial melting may be sufficient to cause melt to be squeezed from its source.

Melts can stay in place long enough to melt to 20% or even 35%, but rocks are rarely melted in excess of 50%, because eventually the melted rock mass becomes a crystal and melt mush that can then ascend *en masse* as a diapir, which may then cause further decompression melting.

Geochemical implications of partial melting

The degree of partial melting is critical for determining what type of magma is produced. The degree of partial melting required to form a melt can be estimated by considering the relative enrichment of incompatible elements versus compatible elements. Incompatible elements commonly include potassium, barium, caesium, rubidium.

Rock types produced by small degrees of partial melting in the Earth's mantle are typically alkaline (Ca, Na), potassic (K) and/or peralkaline (high aluminium to silica ratio). Typically, primitive melts of this composition form lamprophyre, lamproite, kimberlite and sometimes nepheline-bearing mafic rocks such as alkali basalts and essexite gabbros or even carbonatite.

Pegmatite may be produced by low degrees of partial melting of the crust. Some granite-composition magmas are eutectic (or cotectic) melts, and they may be produced by low to high degrees of partial melting of the crust, as well as by fractional crystallization. At high degrees of partial melting of the crust, granitoids such as tonalite, granodiorite and monzonite can be produced, but other mechanisms are typically important in producing them.

Evolution of magmas

Primary melts

When a rock melts, the liquid as a *primary melt*. Primary melts have not undergone any differentiation and represent the starting composition of a magma. In nature it is rare to find primary melts. The leucosomes of migmatites are examples of primary melts. Primary melts derived from the mantle are especially important, and are known as *primitive melts* or primitive magmas. By finding the primitive magma composition of a magma series it is possible to model the composition of the mantle from which a melt was formed, which is important in understanding evolution of the mantle.

Parental melts

Where it is impossible to find the primitive or primary magma composition, it is often useful to attempt to identify a parental melt. A parental melt is a magma composition

from which the observed range of magma chemistries has been derived by the processes of igneous differentiation. It need not be a primitive melt.

For instance, a series of basalt flows are assumed to be related to one another. A composition from which they could reasonably be produced by fractional crystallization is termed a *parental melt*. Fractional crystallization models would be produced to test the hypothesis that they share a common parental melt.

At high degrees of partial melting of the mantle, komatiite and picrite are produced.

Migration

Magma exists and has existed in earth's interior since the formation of the earth. Magma rises toward the Earth's surface as long as it is less dense than the surrounding rock. Once magma stops rising, it can collect in areas called magma chambers. Magma can remain in a chamber until it cools, forming igneous rock, it erupts or moves on to another magma chamber.

Porous rock

When rock is first melted the liquid forms pores in the source rock. Magma in this kind of porous media is very primitive. As the host rock compacts the magma is expelled forming a network that collects magma into magma chambers.

Magma chambers
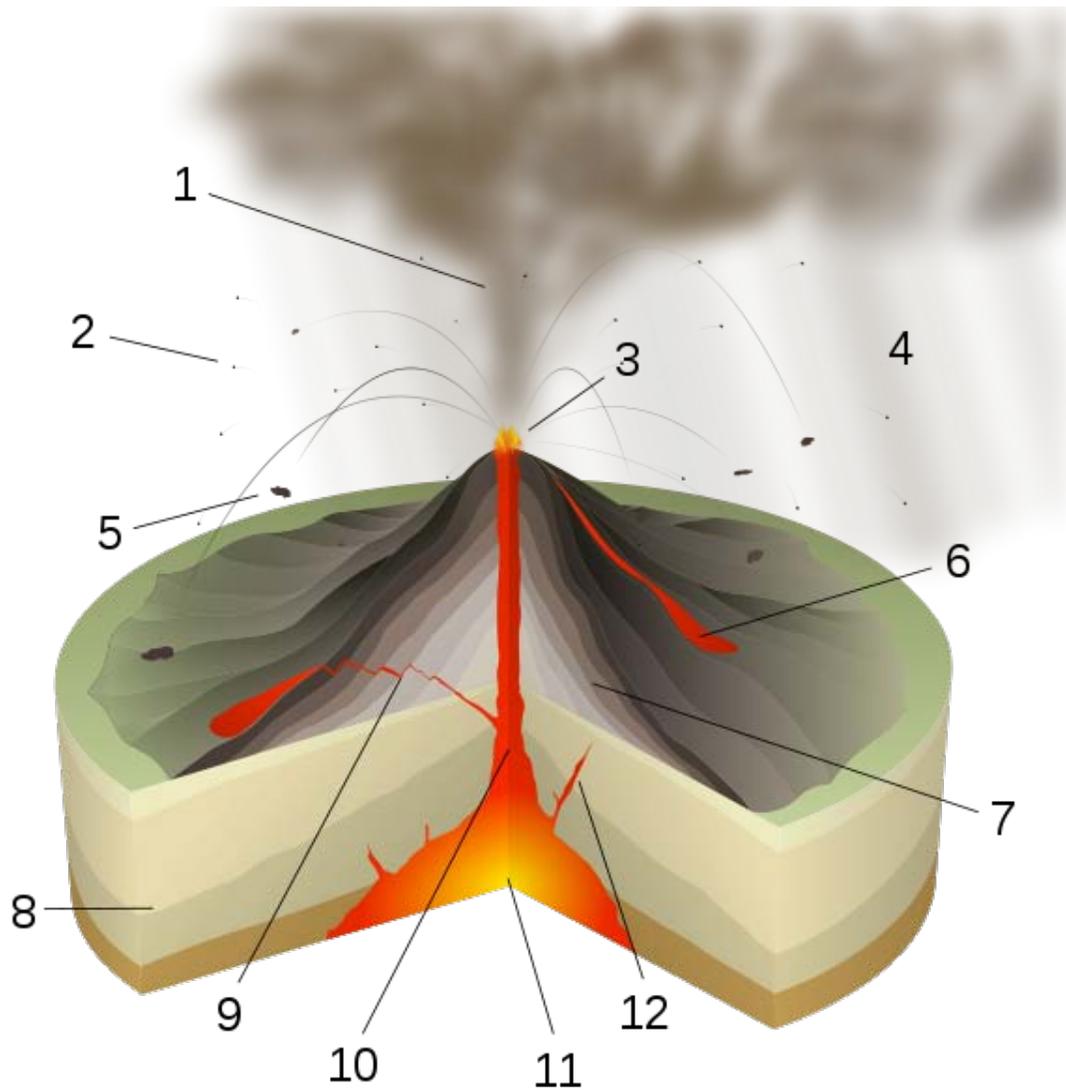
11 - magma chamber

A **magma chamber** is a large underground pool of molten rock found beneath the surface of the Earth. The molten rock in such a chamber is under great pressure, and given enough time, that pressure can gradually fracture the rock around it creating outlets for the magma. If it finds a way to the surface, then the result will be a volcanic eruption; consequently many volcanoes are situated over magma chambers.

Magma chambers are hard to detect, and most of the known ones are therefore close to the surface of the Earth, commonly between 1 km and 10 km under the surface. In geological terms this is extremely close to the surface, although in human terms it is considerably deep underground.

Dynamics of magma chambers

Magma rises through fractures from beneath the crust because it is less dense than the surrounding rock. When the magma cannot find a path upwards it pools into a magma chamber. As more magma rises up below it, the pressure in the chamber grows.

If magma resides in a chamber for a long period, then it can become stratified with lower density components rising to the top and denser materials sinking. It can also start to cool, with the higher melting point components such as olivine crystallising out of the solution, particularly near to the cooler walls of the chamber, and forming a denser conglomerate of minerals which sinks. Any subsequent eruption may produce distinctly layered deposits, for example the deposits from the 79 AD eruption of Mount Vesuvius include a thick layer of white pumice from the upper portion of the magma chamber overlayed with a similar layer of grey pumice produced from material erupted later from lower down in the chamber.

Another effect of the cooling of the chamber is that the solidifying crystals will release the gas (primarily steam) previously dissolved when they were liquid, causing the pressure in the chamber to rise, possibly sufficiently to produce an eruption. Additionally, the removal of the lower melting point components will tend to make the magma more viscous (by increasing the concentration of silicates). Thus, stratification of a magma chamber may result in an increase in the amount of gas within the magma near the top of the chamber, and also make this magma more viscous; potentially leading to a more explosive eruption than would be the case had the chamber not become stratified.

If the magma is not vented to the surface in a volcanic eruption it will slowly cool and crystallize at depth to form an intrusive igneous body composed of granite or gabbro.

Often, a volcano may have a deep magma chamber many kilometres down, which supplies a shallower chamber near the summit. The location of magma chambers can be mapped using seismology: seismic waves from earthquakes move more slowly through liquid rock than solid, allowing measurements to pinpoint the regions of slow movement which identify magma chambers.

As a volcano erupts, emptying the magma chamber, the surrounding rock will collapse into it. If a large amount of magma is erupted, causing the chamber to reduce considerably in volume, then this can result in a depression at the surface called a caldera.

Dike (geology)

Banded gneiss with dike of granite orthogneiss

An intrusion (Notch Peak monzonite) inter-fingers (partly as a dike) with highly-metamorphosed host rock (Cambrian carbonate rocks). From near Notch Peak, House Range, Utah.

A **dike** or **dyke** in geology is a type of sheet intrusion referring to any geologic body that cuts *discordantly* across

- planar wall rock structures, such as bedding or foliation
- massive rock formations, like igneous/magmatic intrusions and salt diapirs.

Dikes can therefore be either intrusive or sedimentary in origin.

Magmatic dikes

An intrusion (Notch Peak monzonite) inter-fingers (partly as a dike) with highly-metamorphosed host rock (Cambrian carbonate rocks). From near Notch Peak, House Range, Utah.

A diabase dike crosscutting horizontal limestone beds in Arizona

A small dike on the Baranof Cross-Island Trail, Alaska

Dikes in the Black Canyon of the Gunnison National Park, Colorado, USA

An intrusive dike is an igneous body with a very high aspect ratio, which means that its thickness is usually much smaller than the other two dimensions. Thickness can vary from sub-centimeter scale to many meters, and the lateral dimensions can extend over many kilometers. A dike is an intrusion into an opening cross-cutting fissure, shouldering aside other pre-existing layers or bodies of rock; this implies that a dike is always younger than the rocks that contain it. Dikes are usually high angle to near vertical in orientation, but subsequent tectonic deformation may rotate the sequence of strata through which the dike propagates so that the latter becomes horizontal. Near horizontal, or conformable intrusions, along bedding planes between strata are called intrusive sills.

Sometimes dikes appear as swarms, consisting of several to hundreds of dikes emplaced more or less contemporaneously during a single intrusive event. The world's largest dike swarm is the Mackenzie dike swarm in the Northwest Territories, Canada.

Shiprock, New Mexico, the volcanic neck in the distance, with radiating dike on its south side. Photo credit: USGS Digital Data Series

Dikes often form as either radial or concentric swarms around plutonic intrusives, volcanic necks or feeder vents in volcanic cones. The latter are known as ring dikes.

Dikes can vary in texture and their composition can range from diabase or basaltic to granitic or rhyolitic, but on a global perspective the basaltic composition prevails, manifesting ascent of vast volumes of mantle-derived magmas through fractured lithosphere throughout Earth history. Pegmatite dikes are extremely coarse crystalline granitic rocks often associated with late-stage granite intrusions or metamorphic segregations. Aplite dikes are fine grained or sugary textured intrusives of granitic composition.

Sedimentary dikes

Clastic dike (left of notebook) in the Chinle Formation in the Island In the Sky District of Canyonlands National Park, Utah.

Sedimentary dikes or clastic dikes are vertical bodies of sedimentary rock that cut off other rock layers. They can form in two ways:

- When a shallow unconsolidated sediment is composed of alternating coarse grained and impermeable clay layers the fluid pressure inside the coarser layers may reach a critical value due to lithostatic overburden. Driven by the fluid pressure the sediment breaks through overlying layers and forms a dike.

- When a soil is under permafrost conditions the pore water is totally frozen. When cracks are formed in such rocks, they may fill up with sediments that fall in from above. The result is a vertical body of sediment that cuts through horizontal layers: a dike.

Volcanic necks

Volcanic plug near Rhumsiki, Far North Region, Cameroon

A **volcanic plug**, also called a **volcanic neck** or **lava neck**, is a volcanic landform created when magma hardens within a vent on an active volcano. When forming, a plug can cause an extreme build-up of pressure if volatile-charged magma is trapped beneath it, and this can sometimes lead to an explosive eruption. If a plug is preserved, erosion may remove the surrounding rock while the erosion-resistant plug remains, producing a distinctive upstanding landform. Examples include Shiprock, New Mexico; The Nut, Australia; and the Pitons of Saint Lucia.

Edinburgh Castle in Scotland is built upon an ancient volcanic plug

Glacial erosion can lead to exposure of the plug on one side, while a long slope of material remains on the lee side. Such landforms are called crag and tail. An example is the Castle Rock in Edinburgh, Scotland.

Volcanic plug in the Ochoco National Forest of Oregon, USA

Further examples of volcanic plugs in the United States include Morro Rock, California; Lizard Head, Colorado; and Laurel Hill, New Jersey. Devils Tower in Wyoming, as well as Little Devils Postpile located in Yosemite National Park, are also thought to be a volcanic plug by many geologists. The only example of a volcanic plug in the eastern USA is the highly eroded Stark's Knob basaltic structure located along the Hudson River near Saratoga Springs, New York. However, some geologists believe Stark's Knob is not a plug at all, but merely an outcrop of an ancient submarine lava flow.

The Ailsa Craig, Bass Rock, North Berwick Law and Dumgoyne hill are examples of volcanic plugs located in Scotland. Borgarvirki is a volcanic plug located in north Iceland. A volcanic plug is situated in the town of Motta Sant'Anastasia in Italy.

New Zealand's Lion Rock was fortified as a refuge for local Maori

There are several volcanic plugs in New Zealand, including the Pinnacles in the Coromandel Peninsula, Bream Head, Paritutu and adjacent Sugar Loaf Islands, and Piha's Lion Rock, which hosted a fortified Maori pa. Australia's Mount Warning is a volcanic plug.

Sigiriya Rock Fortress, Sri Lanka

Another example is Sigiriya (Lion's rock), the hardened magma plug from an extinct and long-eroded volcano. The rock rises 370m and is sheer on all sides, in many places overhanging the base. It hosts an ancient rock fortress and ruins of a castle, one of the seven World Heritage Sites in Sri Lanka.

Trosky Castle ("Panna" Tower), Czech Republic

Petit Piton in Saint Lucia, part of a UNESCO World Heritage site; the mountain rises to a height of nearly 2500 ft.

Cooling of magmas

Although now covered by forest the mountains around Panguipulli Lake in Chile were once magma that crystallized deep in earth's crust, now these magmatic bodies form the Panguipulli Batholith.

There are two known processes by which magma ceases to exist, by volcanic eruption or by plutonism. In both cases the bulk of the magma eventually cools and form igneous rocks.

When magma cools it begins to form solid mineral phases, some of them settles at the bottom of the magma chamber forming cumulates that might form mafic layered intrusions. Magma that cools down slowly in a magma chamber usually ends up as forming bodies of plutonic rocks like gabbro, diorite and granite depending on the comoposition of the magma, while if the magma is erupted it forms volcanic rocks such as basalt, andesite and rhyolite (the extrusive equivalents of gabbro, diorite and granite respectively).

Pluton

A Jurassic pluton of pink monzonite intruded below and beneath a section of gray sedimentary rocks and then was subsequently uplifted and exposed, near Notch Peak, House Range, Utah.

A **pluton** in geology is an intrusive igneous rock (called a **plutonic rock**) body that crystallized from magma slowly cooling below the surface of the Earth. Plutons include batholiths, dikes, sills, laccoliths, lopoliths, and other igneous bodies. In practice, "pluton" usually refers to a distinctive mass of igneous rock, typically kilometers in dimension, without a tabular shape like those of dikes and sills. Batholiths commonly are aggregations of plutons. Examples of plutons include Cardinal Peak and Mount Kinabalu.

The most common rock types in plutons are granite, granodiorite, tonalite, monzonite, and quartz diorite. The term granitoid is used for a general, light colored, coarse-grained igneous rock in which a proper, or more specific name, is not known. Use of granitoid should be restricted to the field wherever possible.

The term originated from *Pluto*, the ancient Roman god of the underworld. The use of the name and concept goes back to the beginnings of the science of geology in the late 18th century and the then hotly debated theories of plutonism (or vulcanism), and neptunism regarding the origin of basalt.

Cumulate Rock

Cumulate rocks are igneous rocks formed by the accumulation of crystals from a magma either by settling or floating. Cumulate rocks are named according to their texture; cumulate texture is diagnostic of the conditions of formation of this group of igneous rocks.

Formation

Cumulate rocks are the typical product of precipitation of solid crystals from a fractionating magma chamber. These accumulations typically occur on the floor of the magma chamber, although they are possible on the roofs if anorthite plagioclase is able to float free of a denser mafic melt.

Cumulates are typically found in ultramafic intrusions, in the base of large ultramafic lava tubes in komatiite and magnesium rich basalt flows and also in some granitic intrusions.

Terminology

Cumulates are named according to their dominant mineralogy and the percentage of crystals to their groundmass.

- *Adcumulates* are rocks containing ~100-93% accumulated magmatic crystals in a fine grained groundmass.
- *Mesocumulates* are rocks with between 93-85% accumulated minerals in a groundmass.
- *Orthocumulates* are rocks containing between 85-75% accumulated minerals in groundmass.

Cumulate rocks are typically named according to the cumulate minerals in order of abundance, and then cumulate type (adcumulate, mesocumulate, orthocumulate), and then accessory or minor phases. For example:

- a layer with 50% plagioclase, 40% pyroxene, 5% olivine and 5% groundmass (in essence a gabbro) would be termed a *plagioclase-pyroxene Adcumulate* with accessory olivine.
- a rock consisting of 80% olivine, 5% magnetite and 15% groundmass is an *olivine mesocumulate*, (in essence a peridotite).

Cumulate terminology is appropriate for use when describing cumulate rocks. In intrusions which have a uniform composition and minimal textural and mineralogical layering or visible crystal accumulations it is inappropriate to describe them according to this convention.

Geochemistry

Cumulate rocks, because they are fractionates of a parental magma, should not be used to infer the composition of a magma from which they are formed. The chemistry of the cumulate itself can inform on the residual melt composition, but several factors need to be considered.

Cumulate chemistry

The chemistry of a cumulate can inform upon the temperature, pressure and chemistry of the melt from which it was formed, but the number of minerals which co-precipitate need to be known, as does the chemistry or mineral species of the precipitated minerals. This is best illustrated by an example;

As an example, a magma of basalt composition that is precipitating cumulates of anorthite plagioclase plus enstatite pyroxene is changing composition by the removal of the elements which make up the precipitated minerals. In this example, the precipitation of anorthite (a calcium aluminium feldspar) removes calcium from the melt, which becomes more depleted in calcium. Enstatite being precipitated from the melt will remove magnesium, so the melt becomes depleted in these elements. This tends to enrich the concentration of other elements - typically sodium, potassium, titanium and iron.

The rock that is made of the accumulated minerals will not have the same composition as the magma. In the above example, the cumulate of anorthite + enstatite is rich in calcium and magnesium, and the melt is depleted in calcium and magnesium. The cumulate rock is a plagioclase-pyroxene cumulate (a gabbro) and the melt is now more felsic and aluminous in composition (trending towards andesite compositions).

In the above example, the plagioclase and pyroxene need not be pure end-member compositions (anorthite-enstatite), and thus the effect of depletion of elements can be complex. The minerals can be precipitated in any ratio within the cumulate; such cumulates can be 90% plagioclase:10% enstatite, through to 10% plagiclase:90% enstatite and remain a gabbro. This also alters the chemistry of the cumulate, and the depletions of the residual melt.

It can be seen that the effect on the composition of the residual melt left behind by the formation of the cumulate is dependent on the composition of the minerals which precipitate, the number of minerals which co-precipitate at the same time, and the ratio of the minerals which co-precipitate. In nature, cumulates usually form from 2 mineral species, although ranges from 1 to 4 mineral species are known. Cumulate rocks which

are formed from one mineral alone are often named after the mineral, for example a 99% magnetite cumulate is known as a magnetitite.

A specific example is the Skaergaard intrusion in Greenland. At Skaergaard a 2500 m thick layered intrusion shows distinct chemical and mineralogic layering:

- plagioclase varies from An_{66} near the base to An_{30} near the top (An_{xx} = anorthite percentage)
  - CaO 10.5 % base to 5.1% top; $Na_2O + K_2O$ 2.3% base to 5.9% top
- olivine varies from Fo_{57} near the base to Fo_0 at the top (Fo_{xx} = forsterite percentage of the olivine)
  - MgO 11.6% Lower zone to 1.7% upper zone; FeO 9.3% lower zone to 22.7% upper zone

The Skaergaard is interpreted to have crystallised from a single confined magma chamber.

Residual melt chemistry

One way to infer the composition of the magma that created the cumulate rocks is to measure groundmass chemistry, but that chemistry is problematic or impossible to sample. Otherwise, complex calculations of averaging cumulate layers must be utilised, which is a complex process. Alternatively, the magma composition can be estimated by assuming certain conditions of magma chemistry and testing them on phase diagrams using measured mineral chemistry. These methods work fairly well for cumulates formed in volcanic conditions (ie; komatiites). Investigating magma conditions of large layered ultramafic intrusions is more fraught with problems.

These methods have their drawbacks, primarily that they must all make certain assumptions which rarely hold true in nature. The foremost problem is the fact that in large ultramafic intrusions, assimilation of wall rocks tends to alter the chemistry of the melt as time progresses, so measuring groundmass compositions may fall short. Mass balance calculations will show deviations from expected ranges, which may infer assimilation has occurred, but then further chemistry must be embarked upon to quantify these findings.

Secondly, large ultramafic intrusions are rarely sealed systems and may be subject to regular injections of fresh, primitive magma, or to loss of volume due to further upward migration of the magma (possibly to feed volcanic vents or dyke swarms). In such cases, calculating magma chemistries may resolve nothing more than the presence of these two processes having affected the intrusion.

Economic importance

The economic importance of cumulate rocks is best represented by three classes of mineral deposits found in ultramafic to mafic layered intrusions.

- Silicate mineral cumulates
- Oxide mineral cumulates
- Sulfide melt cumulates

Silicate mineral cumulates

Silicate minerals are rarely sufficiently valuable to warrant extraction as ore. However, some anorthosite intrusions contain such pure anorthite concentrations that they are mined for feldspar, for use in refractories, glassmaking and other sundry uses (toothpaste, cosmetics, etcetera).

Oxide mineral cumulates

Oxide mineral cumulates form in layered intrusions when fractional crystallisation has progressed enough to allow the crystallisation of oxide minerals which are invariably a form of spinel. This can happen due to fractional enrichment of the melt in iron, titanium or chromium.

These conditions are created by the high-temperature fractionation of highly magnesian olivine and/or pyroxene, which causes a relative iron-enrichment in the residual melt. When the iron content of the melt is sufficiently high enough, magnetite or ilmenite crystallise and, due to their high density, form cumulate rocks. Chromite is generally formed during pyroxene fractionation at low pressures, where chromium is rejected from the pyroxene crystals.

These oxide layers form laterally continuous deposits of rocks containing in excess of 50% oxide minerals. When oxide minerals exceed 90% of the bulk of the interval the rock may be classified according to the oxide mineral, for example *magnetitite*, *ilmenitite* or *chromitite*. Strictly speaking, these would be magnetite orthocumulate, ilmenite orthocumulate and chromite orthocumulates.

Sulfide mineral segregations

Sulfide mineral cumulates in layered intrusions are an important source of nickel, copper, platinum group elements and cobalt. Deposits of a mixed massive or mixed sulfide-silicate 'matrix' of pentlandite, chalcopyrite, pyrrhotite and/or pyrite are formed, occasionally with cobaltite and platinum-tellurium sulfides. These deposits are formed by melt immiscibility between sulfide and silicate melts in a sulfur-saturated magma.

They are not strictly a cumulate rock, as the sulfide is not precipitated as a solid mineral, but rather as immiscible sulfide liquid. However, they are formed by the same processes and accumulate due to their high specific gravity, and can form laterally extensive sulfide 'reefs'. The sulfide minerals generally form an interstitial matrix to a silicate cumulate.

Sulfide mineral segregations can only be formed when a magma attains sulfur saturation. In mafic and ultramafic rocks they form economic Ni, Cu and PGE deposits because

these elements are chalcophile and are strongly partitioned into the sulfide melt. In rare cases, felsic rocks become sulfur saturated and form sulfide segregations. In this case, the typical result is a disseminated form of sulfide mineral, usually a mixture of pyrrhotite, pyrite and chalcopyrite, forming Cu mineralisation. It is very rare but not unknown to see cumulate sulfide rocks in granitic intrusions.

Volcanism

During a volcanic eruption the magma that leaves the underground is called lava (an old word for magma is "subterranean lava"). Lava cools down and solidifies relatively quickly compared to underground bodies of magma. This fast cooling does not allow new crystals to grow large, and a part of the melt does not crystallize at all, becoming glass (obsidian).

Before and during volcanic eruptions, fluids like CO_2 and H_2O partially leave the melt through a process known as exsolution. Magma with low water content becomes increasingly viscous. If massive exsolution occurs when magma heads upwards during a volcanic eruption, the resulting eruption is usually explosive (at least in its initial phases).

Composition, melt structure and properties

Silicate melts are composed mainly of silicon, oxygen, aluminium, alkalis (sodium, potassium, calcium), magnesium and iron. Silicon atoms are in tetrahedral coordination with oxygen, as in almost all silicate minerals, but in melts atomic order is preserved only over short distances. The physical behaviours of melts depend upon their atomic structures as well as upon temperature and pressure and composition.

Viscosity is a key melt property in understanding the behaviour of magmas. More silica-rich melts are typically more polymerized, with more linkage of silica tetrahedra, and so are more viscous. Dissolution of water drastically reduces melt viscosity. Higher-temperature melts are less viscous.

Generally speaking, more mafic magmas, such as those that form basalt, are hotter and less viscous than more silica-rich magmas, such as those that form rhyolite. Low viscosity leads to gentler, less explosive eruptions.

Characteristics of several different magma types are as follows:

> Ultramafic (picritic)
> $SiO_2 < 45\%$
> Fe-Mg >8% up to 32%MgO
> Temperature: up to 1500°C
> Viscosity: Very Low
> Eruptive behavior: gentle or very explosive (kimberilites)

Distribution: divergent plate boundaries, hot spots, convergent plate boundaries; komatiite and other ultramafic lavas are mostly Archean and were formed from a higher geothermal gradient and are unknown in the present

Mafic (basaltic)

SiO_2 < 50%

FeO and MgO typically < 10 wt%

Temperature: up to ~1300°C

Viscosity: Low

Eruptive behavior: gentle

Distribution: divergent plate boundaries, hot spots, convergent plate boundaries

Intermediate (andesitic)

SiO_2 ~ 60%

Fe-Mg: ~ 3%

Temperature: ~1000°C

Viscosity: Intermediate

Eruptive behavior: explosive or effusive

Distribution: convergent plate boundaries, island arcs

Felsic (rhyolitic)

SiO_2 >70%

Fe-Mg: ~ 2%

Temp: < 900°C

Viscosity: High

Eruptive behavior: explosive or effusive

Distribution: hot spots in continental crust (Yellowstone National Park), continental rifts

Temperature

At any given pressure and for any given composition of rock, a rise in temperature past the solidus will cause melting. Within the solid earth, the temperature of a rock is controlled by the geothermal gradient and the radioactive decay within the rock. The geothermal gradient averages about 25 °C/km with a wide range from a low of 5-10 °C/km within oceanic trenches and subduction zones to 30-80 °C/km under mid-ocean ridges and volcanic arc environments.

Pressure

As magma buoyantly rises it will cross the solidus-liquidus and its temperature will reduce by adiabatic cooling. At this point it will liquefy and if erupted onto the surface will form lava. Melting can also occur due to a reduction in pressure by a process known as decompression melting.

Composition

It is usually very difficult to change the bulk composition of a large mass of rock, so composition is the basic control on whether a rock will melt at any given temperature and

pressure. The composition of a rock may also be considered to include *volatile* phases such as water and carbon dioxide.

The presence of volatile phases in a rock under pressure can stabilize a melt fraction. The presence of even 0.8% water may reduce the temperature of melting by as much as 100 °C. Conversely, the loss of water and volatiles from a magma may cause it to essentially freeze or solidify.

Also a major portion of all magma is silica, which is a compound of silicon and oxygen. Magma also contains gases, which expand as the magma rises. Magma that is high in silica resists flowing, so expanding gases are trapped in it. Pressure builds up until the gases blast out in a violent, dangerous explosion. Magma that is relatively poor in silica flows easily, so gas bubbles move up through it and escape fairly gently. Though an eruption of silica-poor magma can throw lava high into the air, forming lava fountains, visitors can usually watch safely nearby.

Chapter- 4

Mantle

The **mantle** is a part of a terrestrial planet or other rocky body large enough to have differentiated by density. The interior of the Earth, similar to the other terrestrial planets, is chemically divided into layers. The mantle is a highly viscous layer between the crust and the outer core. Earth's mantle is a rocky shell about 2,890 km (1,800 mi) thick that constitutes about 90 percent of Earth's volume. It is predominantly solid and encloses the iron-rich hot core, which occupies about 15 percent of Earth's volume. Past episodes of melting and volcanism at the shallower levels of the mantle have produced a thin crust of crystallized melt products near the surface, upon which we live. The gases evolved during the melting of Earth's mantle have a large effect on the composition and abundance of Earth's atmosphere. Information about structure and composition of the mantle either result from geophysical investigation or from direct geoscientific analyses on Earth mantle derived xenoliths.

Two main zones are distinguished in the upper mantle: the inner asthenosphere composed of plastic flowing rock, about 200 km thick, and the lowermost part of the lithosphere, composed of rigid rock, about 50 to 120 km thick. A thin crust, the upper part of the lithosphere, surrounds the mantle and is about 5 to 75 km thick.
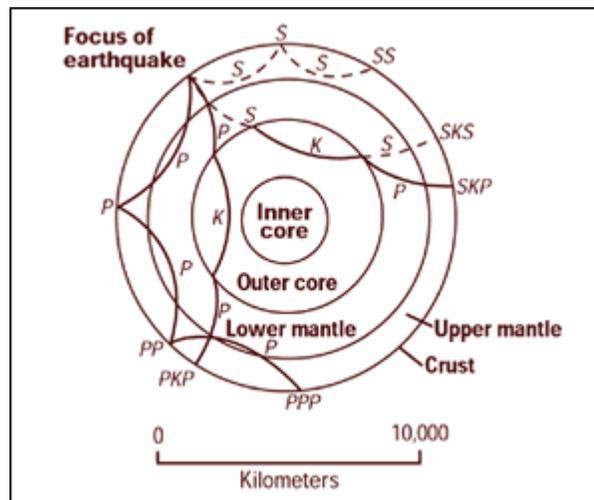
Structure

The mantle is divided into sections based upon results from seismology. These layers (and their depths) are the following: the upper mantle (starting at the Moho, or base of the crust around 7 to 35 km, downward to 410 km), the transition zone (410–660 km), the lower mantle (660–2891 km), and in the bottom of the latter region there is the anomalous D" layer with a variable thickness (on average ~200 km thick).

The top of the mantle is defined by a sudden increase in seismic velocity, which was first noted by Andrija Mohorovičić in 1909; this boundary is now referred to as the "Mohorovičić discontinuity" or "Moho". The uppermost mantle plus overlying crust are relatively rigid and form the lithosphere, an irregular layer with a maximum thickness of perhaps 200 km. Below the lithosphere the upper mantle becomes notably more plastic in its rheology. In some regions below the lithosphere, the seismic velocity is reduced; this so-called low-velocity zone (LVZ) extends down to a depth of several hundred km. Inge Lehmann discovered a seismic discontinuity at about 220 km depth; although this discontinuity has been found in other studies, it is not known whether the discontinuity is

ubiquitous. The transition zone is an area of great complexity; it physically separates the upper and lower mantle. Very little is known about the lower mantle apart from that it appears to be relatively seismically homogeneous. The D" layer at the core–mantle boundary separates the mantle from the core.

Characteristics

The mantle differs substantially from the crust in its mechanical characteristics and its chemical composition. The distinction between crust and mantle is based on chemistry, rock types, rheology and seismic characteristics. The crust is, in fact, a product of mantle melting. Partial melting of mantle material is believed to cause incompatible elements to separate from the mantle rock, with less dense material floating upward through pore spaces, cracks, or fissures, to cool and freeze at the surface. Typical mantle rocks have a higher magnesium to iron ratio, and a smaller proportion of silicon and aluminium than the crust. This behavior is also predicted by experiments that partly melt rocks thought to be representative of Earth's mantle.

Mapping the interior of the Earth with earthquake waves

Mantle rocks shallower than about 410 km depth consist mostly of olivine, pyroxenes, spinel-structure minerals, and garnet; typical rock types are thought to be peridotite, dunite (olivine-rich peridotite), and eclogite. Between about 400 km and 650 km depth, olivine is not stable and is replaced by high pressure polymorphs with approximately the same composition: one polymorph is wadsleyite (also called *beta-spinel* type), and the other is ringwoodite (a mineral with the *gamma-spinel* structure). Below about 650 km, all of the minerals of the upper mantle begin to become unstable. The most abundant minerals present have structures (but not compositions) like that of the mineral perovskite followed by the magnesium/iron oxide ferropericlase. The changes in mineralogy at about 400 and 650 km yield distinctive signatures in seismic records of the Earth's interior, and like the moho, are readily detected using seismic waves. These changes in mineralogy may influence mantle convection, as they result in density changes and they

may absorb or release latent heat as well as depress or elevate the depth of the polymorphic phase transitions for regions of different temperatures. The changes in mineralogy with depth have been investigated by laboratory experiments that duplicate high mantle pressures, such as those using the diamond anvil.

Composition of Earth's mantle in weight percent			
Element	Amount	Compound	Amount
O	44.8		
Si	21.5	SiO_2	46
Mg	22.8	MgO	37.8
Fe	5.8	FeO	7.5
Al	2.2	Al_2O_3	4.2
Ca	2.3	CaO	3.2
Na	0.3	Na_2O	0.4
K	0.03	K_2O	0.04
Sum	99.7	Sum	99.1

The inner core is solid, the outer core is liquid, and the mantle solid/plastic. This is because of the relative melting points of the different layers (nickel-iron core, silicate crust and mantle) and the increase in temperature and pressure as one moves deeper into the Earth. At the surface both nickel-iron alloys and silicates are sufficiently cool to be solid. In the upper mantle, the silicates are generally solid (localised regions with small amounts of melt exist); however, as the upper mantle is both hot and under relatively little pressure, the rock in the upper mantle has a relatively low viscosity, i.e. it is relatively fluid. In contrast, the lower mantle is under tremendous pressure and therefore has a higher viscosity than the upper mantle. The metallic nickel-iron outer core is liquid despite the enormous pressure as it has a melting point that is lower than the mantle silicates. The inner core is solid due to the overwhelming pressure found at the center of the planet.

Temperature

In the mantle, temperatures range between 500 to 900 °C (932 to 1,652 °F) at the upper boundary with the crust to over 4,000 °C (7,230 °F) at the boundary with the core. Although the higher temperatures far exceed the melting points of the mantle rocks at the surface (about 1200 °C for representative peridotite), the mantle is almost exclusively solid. The enormous lithostatic pressure exerted on the mantle prevents melting, because the temperature at which melting begins (the solidus) increases with pressure.

Movement
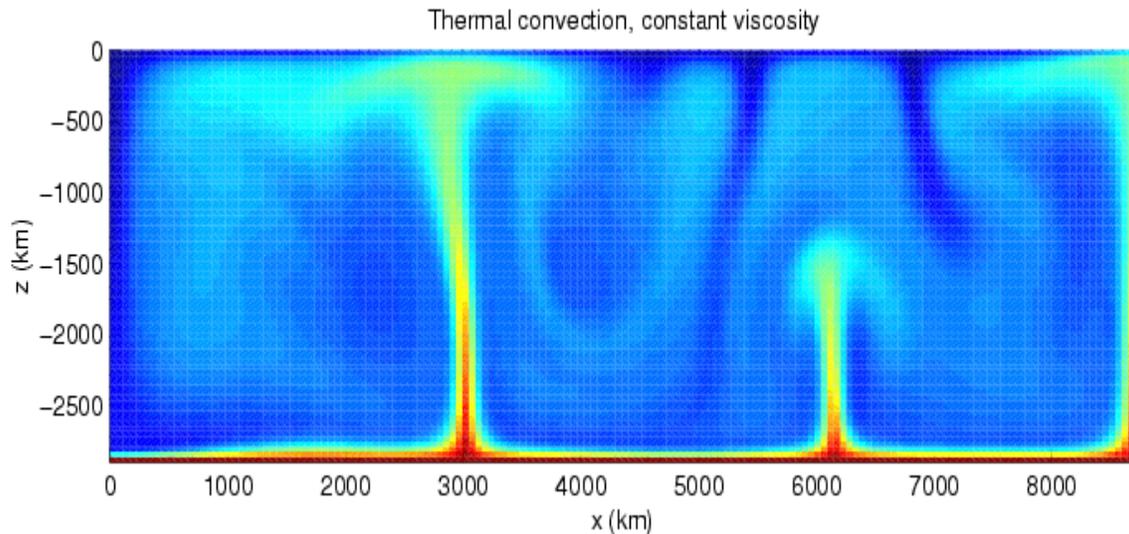
This figure is a snapshot of one time-step in a model of mantle convection. Colors closer to red are hot areas and colors closer to blue are cold areas. In this figure, heat received at the core-mantle boundary results in thermal expansion of the material at the bottom of the model, reducing its density and causing it to send plumes of hot material upwards. Likewise, cooling of material at the surface results in its sinking.

Due to the temperature difference between the Earth's surface and outer core, and the ability of the crystalline rocks at high pressure and temperature to undergo slow, creeping, viscous-like deformation over millions of years, there is a convective material circulation in the mantle. Hot material upwells, while cooler (and heavier) material sinks downward. Downward motion of material often occurs at convergent plate boundaries called subduction zones, while upwelling of material can take the form of plumes. Locations on the surface that lie over plumes will often increase in elevation (due to the buoyancy of the hotter, less-dense plume beneath) and exhibit hot spot volcanism.

The convection of the Earth's mantle is a chaotic process (in the sense of fluid dynamics), which is thought to be an integral part of the motion of plates. Plate motion should not be confused with the older term continental drift which applies purely to the movement of the crustal components of the continents. The movements of the lithosphere and the underlying mantle are coupled since descending lithosphere is an essential component of convection in the mantle. The observed continental drift is a complicated relationship between the forces causing oceanic lithosphere to sink and the movements within Earth's mantle.

Although there is a tendency to larger viscosity at greater depth, this relation is far from linear, and shows layers with dramatically decreased viscosity, in particular in the upper mantle and at the boundary with the core. The mantle within about 200 km above the core-mantle boundary appears to have distinctly different seismic properties than the

mantle at slightly shallower depths; this unusual mantle region just above the core is called **D″** ("D double-prime"), a nomenclature introduced over 50 years ago by the geophysicist Keith Bullen. **D″** may consist of material from subducted slabs that descended and came to rest at the core-mantle boundary and/or from a new mineral polymorph discovered in perovskite called post-perovskite.

Earthquakes at shallow depths are a result of stick-slip faulting, however, below about 50 km the hot, high pressure conditions ought to inhibit further seismicity. The mantle is also considered to be viscous, and so incapable of brittle faulting. However, in subduction zones, earthquakes are observed down to 670 km. A number of mechanisms have been proposed to explain this phenomenon, including dehydration, thermal runaway, and phase change.

The geothermal gradient can be lowered where cool material from the surface sinks downward, increasing the strength of the surrounding mantle, and allowing earthquakes to occur down to a depth of 400 km and 670 km.

The pressure at the bottom of the mantle is ~136 GPa (1.4 million atm). There exists increasing pressure as one travels deeper into the mantle, since the material beneath has to support the weight of all the material above it. The entire mantle, however, is still thought to deform like a fluid on long timescales, with permanent plastic deformation accommodated by the movement of point, line, and/or planar defects through the solid crystals comprising the mantle. Estimates for the viscosity of the upper mantle range between 10^{19} and 10^{24} Pa·s, depending on depth, temperature, composition, state of stress, and numerous other factors. Thus, the upper mantle can only flow very slowly. However, when large forces are applied to the uppermost mantle it can become weaker, and this effect is thought to be important in allowing the formation of tectonic plate boundaries.

Exploration

Exploration of the mantle is generally conducted at the seabed rather than on land due to the relative thinness of the oceanic crust as compared to the significantly thicker continental crust.

The first attempt at mantle exploration, known as Project Mohole, was abandoned in 1966 after repeated failures and cost over-runs. The deepest penetration was approximately 180 m (590 ft). In 2005 the third-deepest oceanic borehole hole reached 1,416 metres (4,646 ft) below the sea floor from the ocean drilling vessel JOIDES Resolution.

On 5 March 2007, a team of scientists on board the RRS *James Cook* embarked on a voyage to an area of the Atlantic seafloor where the mantle lies exposed without any crust covering, mid-way between the Cape Verde Islands and the Caribbean Sea. The exposed site lies approximately three kilometres beneath the ocean surface and covers thousands of square kilometres.

A relatively difficult attempt to retrieve samples from the Earth's mantle was scheduled for later in 2007. As part of the Chikyu Hakken mission, was to use the Japanese vessel 'Chikyu' to drill up to 7,000 m (23,000 ft) below the seabed. This is nearly three times as deep as preceding oceanic drillings.

A novel method of exploring the uppermost few hundred kilometres of the Earth was recently proposed, consisting of a small, dense, heat-generating probe which melts its way down through the crust and mantle while its position and progress are tracked by acoustic signals generated in the rocks. The probe consists of an outer sphere of tungsten about one metre in diameter inside which is a cobalt-60 radioactive heat source. It was calculated that such a probe will reach the oceanic Moho in less than 6 months and attain minimum depths of well over 100 km in a few decades beneath both oceanic and continental lithosphere.

Exploration can also be aided through computer simulations of the evolution of the mantle. In 2009, a supercomputer application provided new insight into the distribution of mineral deposits, especially isotopes of iron, from when the mantle developed 4.5 billion years ago.

Chapter- 5

Plate Tectonics
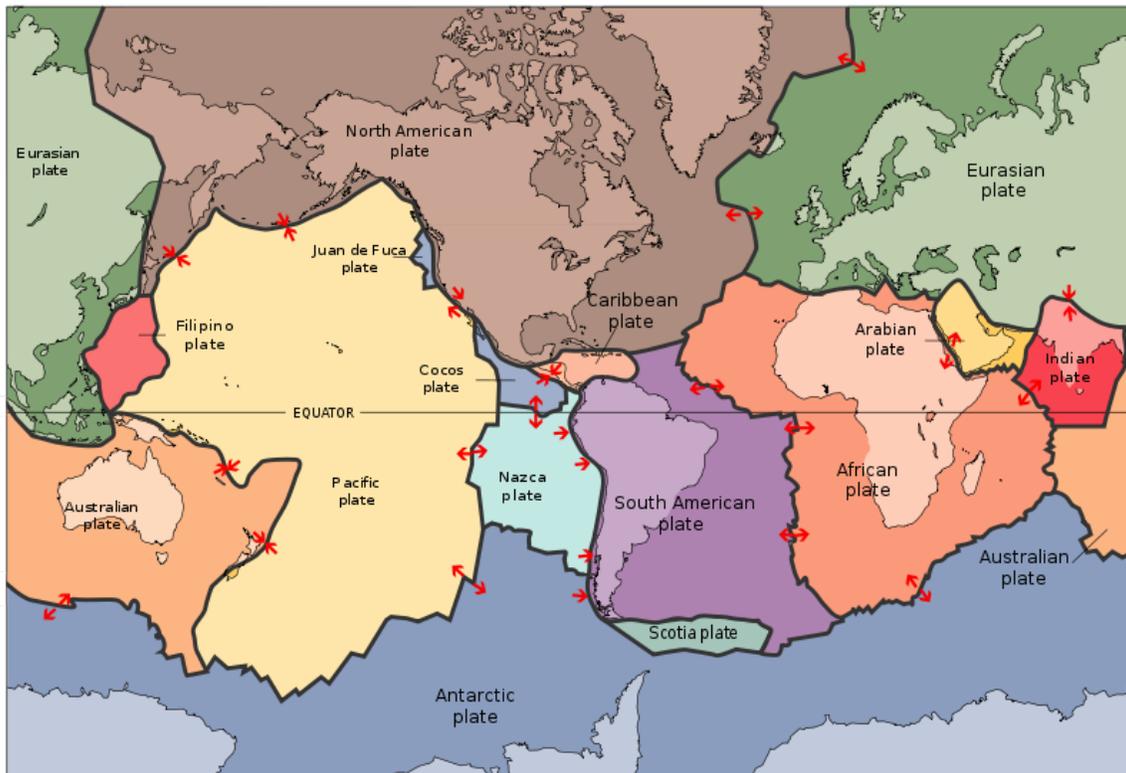
The tectonic plates of the world were mapped in the second half of the 20th century

Plate tectonics (from the Late Latin *tectonicus*, from the Greek: τεκτονικός "pertaining to building") (Little, Fowler & Coulson 1990) is a scientific theory which describes the large scale motions of Earth's lithosphere. The theory builds on the older concepts of continental drift, developed during the first decades of the 20th century (one of the most famous advocates was Alfred Wegener), and was accepted by the majority of the Geoscientific community when the concepts of seafloor spreading were developed in the late 1950s and early 1960s. The lithosphere is broken up into what are called "tectonic plates". In the case of the Earth, there are currently seven to eight major (depending on how they are defined) and many minor plates. The lithospheric plates ride on the

asthenosphere. These plates move in relation to one another at one of three types of plate boundaries: convergent, or collisional boundaries; divergent boundaries, also called spreading centers; and conservative transform boundaries. Earthquakes, volcanic activity, mountain-building, and oceanic trench formation occur along these plate boundaries. The lateral relative movement of the plates varies, though it is typically 0–100 mm annually (Read & Watson 1975).

The tectonic plates are composed of two types of lithosphere: thicker continental and thin oceanic. The upper part is called the crust, again of two types (continental and oceanic). This means that a plate can be of one type, or of both types. One of the main points the theory proposes is that the amount of surface of the (continental and oceanic) plates that disappear in the mantle along the convergent boundaries by subduction is more or less in equilibrium with the new (oceanic) crust that is formed along the divergent margins by seafloor spreading. This is also referred to as the "conveyor belt" principle. In this way, the total surface of the Globe remains the same. This is in contrast with earlier theories advocated before the Plate Tectonics "paradigm", as it is sometimes called, became the main scientific model, theories that proposed gradual shrinking (contraction) or gradual expansion of the Globe, and that still exist in science as alternative models.

Regarding the driving mechanism of the plates various models co-exist: Tectonic plates are able to move because the Earth's lithosphere has a higher strength and lower density than the underlying asthenosphere. Lateral density variations in the mantle result in convection. Their movement is thought to be driven by a combination of the motion of seafloor away from the spreading ridge (due to variations in topography and density of the crust that result in differences in gravitational forces) and drag, downward suction, at the subduction zones. A different explanation lies in different forces generated by the rotation of the Globe and tidal forces of the Sun and the Moon. The relative importance of each of these factors is unclear.

Key principles

The outer layers of the Earth are divided into lithosphere and asthenosphere. This is based on differences in mechanical properties and in the method for the transfer of heat. Mechanically, the lithosphere is cooler and more rigid, while the asthenosphere is hotter and flows more easily. In terms of heat transfer, the lithosphere loses heat by conduction whereas the asthenosphere also transfers heat by convection and has a nearly adiabatic temperature gradient. This division should not be confused with the *chemical* subdivision of these same layers into the mantle (comprising both the asthenosphere and the mantle portion of the lithosphere) and the crust: a given piece of mantle may be part of the lithosphere or the asthenosphere at different times, depending on its temperature and pressure.

The key principle of plate tectonics is that the lithosphere exists as separate and distinct *tectonic plates*, which ride on the fluid-like (visco-elastic solid) asthenosphere. Plate motions range up to a typical 10–40 mm/a (Mid-Atlantic Ridge; about as fast as fingernails grow), to about 160 mm/a (Nazca Plate; about as fast as hair grows) (Zhen

Shao 1997; Hancock, Skinner & Dineley 2000). The driving mechanism behind this movement is described below in a separate section.

Tectonic lithosphere plates consist of lithospheric mantle overlain by either or both of two types of crustal material: oceanic crust (in older texts called *sima* from silicon and magnesium) and continental crust (*sial* from silicon and aluminium). Average oceanic lithosphere is typically 100 km thick (Turcotte & Schubert 2002); its thickness is a function of its age: as time passes, it conductively cools and becomes thicker. Because it is formed at mid-ocean ridges and spreads outwards, its thickness is therefore a function of its distance from the mid-ocean ridge where it was formed. For a typical distance oceanic lithosphere must travel before being subducted, the thickness varies from about 6 km thick at mid-ocean ridges to greater than 100 km at subduction zones; for shorter or longer distances, the subduction zone (and therefore also the mean) thickness becomes smaller or larger, respectively (Turcotte & Schubert 2002). Continental lithosphere is typically ~200 km thick, though this also varies considerably between basins, mountain ranges, and stable cratonic interiors of continents. The two types of crust also differ in thickness, with continental crust being considerably thicker than oceanic (35 km vs. 6 km) (Turcotte & Schubert 2002).

The location where two plates meet is called a *plate boundary*, and plate boundaries are commonly associated with geological events such as earthquakes and the creation of topographic features such as mountains, volcanoes, mid-ocean ridges, and oceanic trenches. The majority of the world's active volcanoes occur along plate boundaries, with the Pacific Plate's Ring of Fire being most active and most widely known. These boundaries are discussed in further detail below.

As explained above, tectonic plates can include continental crust or oceanic crust, and many plates contain both. For example, the African Plate includes the continent and parts of the floor of the Atlantic and Indian Oceans. The distinction between oceanic crust and continental crust is based on their modes of formation. Oceanic crust is formed at sea-floor spreading centers, and continental crust is formed through arc volcanism and accretion of terranes through tectonic processes; though some of these terranes may contain ophiolite sequences, which are pieces of oceanic crust, these are considered part of the continent when they exit the standard cycle of formation and spreading centers and subduction beneath continents. Oceanic crust is also denser than continental crust owing to their different compositions. Oceanic crust is denser because it has less silicon and more heavier elements ("mafic") than continental crust ("felsic") (Schmidt & Harbert 1998). As a result of this density stratification, oceanic crust generally lies below sea level (for example most of the Pacific Plate), while the continental crust buoyantly projects above sea level.

Types of plate boundaries
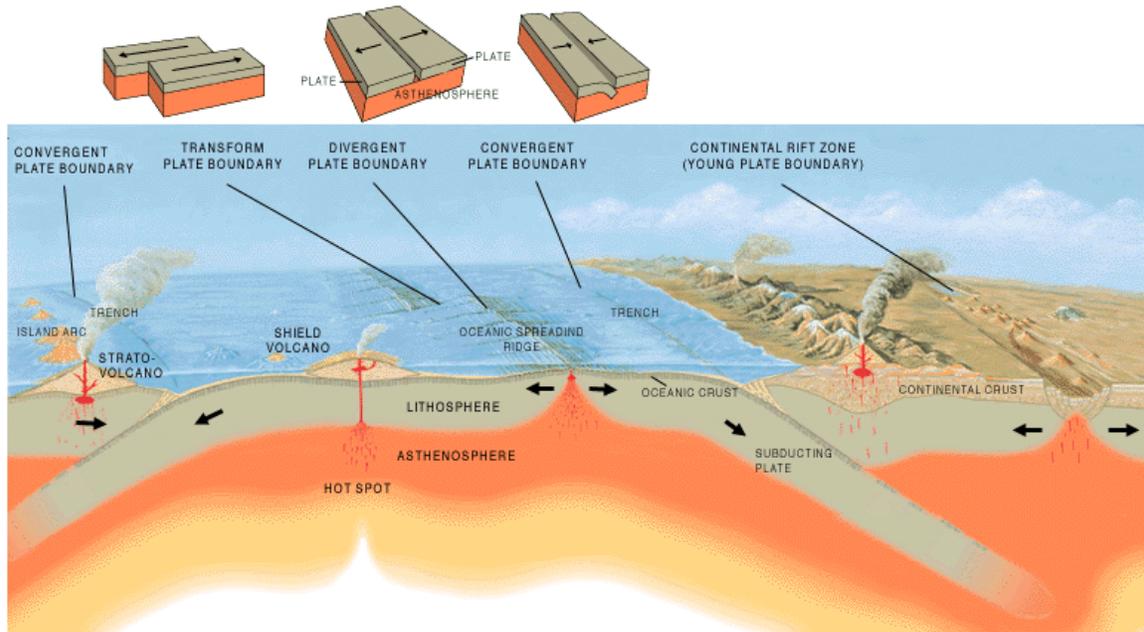
Three types of plate boundary

Basically, three types of plate boundaries exist (Meissner 2002, p. 100), with a fourth, mixed type, characterized by the way the plates move relative to each other. They are associated with different types of surface phenomena. The different types of plate boundaries are:

1. *Transform boundaries (Conservative)* occur where plates slide or, perhaps more accurately, grind past each other along transform faults. The relative motion of the two plates is either sinistral (left side toward the observer) or dextral (right side toward the observer). The San Andreas Fault in California is an example of a transform boundary exhibiting dextral motion.
2. *Divergent boundaries (Constructive)* occur where two plates slide apart from each other. Mid-ocean ridges (e.g., Mid-Atlantic Ridge) and active zones of rifting (such as Africa's Great Rift Valley) are both examples of divergent boundaries.
3. *Convergent boundaries (Destructive)* (or *active margins*) occur where two plates slide towards each other commonly forming either a subduction zone (if one plate moves underneath the other) or a continental collision (if the two plates contain continental crust). Deep marine trenches are typically associated with subduction zones. The subducting slab contains many hydrous minerals, which release their water on heating; this water then causes the mantle to melt, producing volcanism. Examples of this are the Andes mountain range in South America and the Japanese island arc.
4. *Plate boundary zones* occur where the effects of the interactions are unclear and the broad belt boundaries are not well defined.

Driving forces of plate motion

Plate tectonics is basically a kinematic phenomenon: Earth scientists agree upon the observation and deduction that the plates have moved one with respect to the other, and debate and find agreements on how and when. But still a major question remains on what is the motor behind this movement; the geodynamic mechanism, and here science diverges in different theories.

Generally, it is accepted that tectonic plates are able to move because of the relative density of oceanic lithosphere and the relative weakness of the asthenosphere. Dissipation of heat from the mantle is acknowledged to be the original source of energy driving plate tectonics, through convection or large scale upwelling and doming. As a consequence, in the current view, although it is still a matter of some debate, because of the excess density of the oceanic lithosphere sinking in subduction zones a powerful source of plate motion is generated. When the new crust forms at mid-ocean ridges, this oceanic lithosphere is initially less dense than the underlying asthenosphere, but it becomes denser with age, as it conductively cools and thickens. The greater density of old lithosphere relative to the underlying asthenosphere allows it to sink into the deep mantle at subduction zones, providing most of the driving force for plate motions. The weakness of the asthenosphere allows the tectonic plates to move easily towards a subduction zone. Although subduction is believed to be the strongest force driving plate motions, it cannot be the only force since there are plates such as the North American Plate which are moving, yet are nowhere being subducted. The same is true for the enormous Eurasian Plate. The sources of plate motion are a matter of intensive research and discussion among earth scientists. One of the main points is that the kinematic pattern of the movements itself should be separated clearly from the possible geodynamic mechanism that is invoked as the driving force of the observed movements, as some patterns may be explained by more than one mechanism (van Dijk 1992, van Dijk & Okkes 1991). Basically, the driving forces that are advocated at the moment, can be divided in three categories: mantle dynamics related, gravity related (mostly secondary forces), and Earth rotation related.

Mantle dynamics related driving forces

For a considerable period of around 25 years (last quarter of the twentieth century) the leading theory envisaged large scale convection currents in the upper mantle which are transmitted through the asthenosphere as the main driving force of the tectonic plates. This theory was launched by Arthur Holmes and some forerunners in the 1930s and was immediately recognised as the solution for the acceptance of the theory discussed since its occurrence in the papers of Alfred Wegener in the early years of the century. It was, though, long debated because the leading ("fixist") theory was still envisaging a static Earth without moving continents, up until the major break–throughs in the early sixties.

Two– and three–dimensional imaging of the Earth's interior (seismic tomography) shows that there is a laterally varying density distribution throughout the mantle. Such density variations can be material (from rock chemistry), mineral (from variations in mineral structures), or thermal (through thermal expansion and contraction from heat energy).

The manifestation of this varying lateral density is mantle convection from buoyancy forces (Tanimoto & Lay 2000).

How mantle convection relates directly and indirectly to the motion of the plates is a matter of ongoing study and discussion in geodynamics. Somehow, this energy must be transferred to the lithosphere in order for tectonic plates to move. There are essentially two types of forces that are thought to influence plate motion: friction and gravity.

> Basal drag (friction): The plate motion is in this way driven by friction between the convection currents in the asthenosphere and the more rigid overlying floating lithosphere.
> Slab suction (gravity): Local convection currents exert a downward frictional pull on plates in subduction zones at ocean trenches. Slab suction may occur in a geodynamic setting wherein basal tractions continue to act on the plate as it dives into the mantle (although perhaps to a greater extent acting on both the under and upper side of the slab).

Lately, the convection theory is much debated as modern techniques based on 3D seismic tomography of imaging the internal structure of the Earth's mantle still fail to recognise these predicted large scale convection cells. Therefore, alternative patterns of mantle dynamics have been proposed:

> In the theory of plume tectonics developed during the 1990s, a modified concept of mantle convection currents is used, related to super plumes rising from the deeper mantle which would be the drivers or the substitutes of the major convection cells. These ideas, which find their roots in the early 1930s with the so-called "fixistic" ideas of the European and Russian Earth Science Schools, find resonance in the modern theories which envisage hot spots/mantle plumes in the mantle which remain fixed and are overridden by oceanic and continental lithosphere plates during time, and leave their traces in the geological record (though these phenomena are not invoked as real driving mechanisms, but rather as a modulator). The modern theories that continue building on the older mantle doming concepts and see the movements of the plates a secondary phenomena, are beyond the scope of this page and are discussed elsewhere for example on the plume tectonics page.
> Another suggestion is that the mantle flows neither in cells nor large plumes, but rather as a series of channels just below the Earth's crust which then provide basal friction to the lithosphere. This theory is called "surge tectonics" and became quite popular in geophysics and geodynamics during the 1980s and 1990s (Smoot et al. 1996).

Gravity related driving forces

Gravity related forces are usually invoked as secondary phenomena within the framework of a more general driving mechanism such as the various forms of mantle dynamics described above.

Gravitational sliding away from a spreading ridge: According to many authors, plate motion is driven by the higher elevation of plates at ocean ridges. As oceanic lithosphere is formed at spreading ridges from hot mantle material, it gradually cools and thickens with age (and thus distance from the ridge). Cool oceanic lithosphere is significantly denser than the hot mantle material from which it is derived and so with increasing thickness it gradually subsides into the mantle to compensate the greater load. The result is a slight lateral incline with distance from the ridge axis.

This force is regarded as a secondary force often referred to as "ridge-push". This is a misnomer as nothing is "pushing" horizontally and tensional features are dominant along ridges. It is more accurate to refer to this mechanism as gravitational sliding as variable topography across the totality of the plate can vary considerably and the topography of spreading ridges is only the most prominent feature. Other mechanisms generating this gravitational secondary force are for example:

> Flexural bulging of the lithosphere before it dives underneath an adjacent plate, for instance, produces a clear topographical feature that can offset or at least affect the influence of topographical ocean ridges.
> Mantle plumes and hot spots impinging on the underside of tectonic plates can drastically alter the topography of the ocean floor. Some of these, on a larger scale, are seen as the major driving force of the plates (see below).

Slab-pull: Current scientific opinion is that the asthenosphere is insufficiently competent or rigid to directly cause motion by friction along the base of the lithosphere. Slab pull is therefore most widely thought to be the greatest force acting on the plates. In this current understanding, plate motion is mostly driven by the weight of cold, dense plates sinking into the mantle at trenches (Conrad & Lithgow-Bertelloni 2002). Recent models indicate that trench suction plays an important role as well. However, as the North American Plate is nowhere being subducted, yet it is in motion presents a problem. The same holds for the African, Eurasian, and Antarctic plates. Slab pull is especially invoked in areas where remnants of older lithosphere become trapped along convergence zones e.g. as relicts in collisional belts, which, sinking into the mantle and rolling backwards, exert a pull on the overlying crust.

Gravitational sliding away from mantle doming: According to older theories one of the driving mechanisms of the plates is the existence of large scale asthenosphere/mantle domes, which cause the gravitational sliding of lithosphere plates away from them. This gravitational sliding represents a secondary phenomenon of this, basically vertically oriented mechanism. This can act on various scales, from the small scale of one island arc up to the larger scale of an entire ocean basin.

Earth rotation related driving forces

Alfred Wegener, being a meteorologist, had proposed tidal forces and pole flight Force as main driving mechanisms for continental drift. However, these forces were considered far too small to cause continental motion as the concept then was of continents plowing

through oceanic crust. Therefore, also Wegener in his last edition of his book in 1929 converted to convection currents as the main driving force.

In the plate tectonics context (accepted since the seafloor spreading proposals of Heezen, Hess, Dietz, Morley, Vine and Matthews -see below- during the early 1960s), though, oceanic crust in motion *with* the continents which made the proposals related to Earth rotation to be reconsidered, also in more recent literature, these are:

1. Tidal drag due to the gravitational force the Moon (and the Sun) exerts on the crust of the Earth
2. Shear strain of the Earth globe due to N-S compression related to the rotation and modulations of it
3. Pole flight force: equatorial drift due to rotation and centrifugal effects: tendency of the plates to move from the poles to the equator ("*Polflucht*")
4. Coriolis effect the plates suffer when they move around the globe (coriolis effect/law of Buys Ballot)
5. Global deformation of the geoid due to small displacements of rotational pole with respect to the Earth crust
6. Other smaller deformation effects of the crust due to wobbles and spin movements of the Earth rotation on a smaller time scale.

In order for these mechanisms to be overall valid, systematic relationships should exist all over the Globe between the orientation and kinematics of deformation, and the geographical latitudinal and longitudinal grid of the Earth itself. Ironically, these systematic relations studies in the second half of the nineteenth century and the first half of the twentieth century do underline exactly the opposite: that the plates had not moved in time, that the deformation grid was fixed with respect to the Earth equator and axis, and that gravitational driving forces were generally acting vertically and caused only locally horizontal movements (the so-called pre-plate tectonic, "fixist theories"). Later studies (discussed below on this page) therefore invoked many of the relationships recognised during this pre-plate tectonics period, to support their theories.

Of the many forces discussed in this paragraph, tidal force is still highly debated and defended as a possible principle driving force, whereas the other forces are used or in global geodynamic models not using the plate tectonics concepts (therefore beyond the discussions treated in this section), or proposed as minor modulations within the overall plate tectonics model.

In 1973 George W. Moore of the USGS and R. C. Bostrom presented evidence for a general westward drift of the Earth's lithosphere with respect to the mantle, and, therefore, tidal forces or tidal lag or "friction" due to the Earth's rotation and the forces acting upon it by the Moon being a driving force for plate tectonics: as the Earth spins eastward beneath the moon, the moon's gravity ever so slightly pulls the Earth's surface layer back westward, just like proposed by Alfred Wegener (see above). In a more recent 2006 study (Scoppola et al. 2006), scientists rediscussed and advocated these earlier proposed ideas. It has also been suggested recently in Lovett (2006) that this observation

may also explain why Venus and Mars have no plate tectonics, since Venus has no moon and Mars' moons are too small to have significant tidal effects on Mars. In a recent paper by Torsvik et al. (2010) it was suggested that, on the other hand, it can easily be observed that many plates are moving north and eastward, and that the dominantly westward motion of the Pacific ocean basins derives simply from the eastward bias of the Pacific spreading center (which is not a predicted manifestation of such lunar forces). In the same paper the authors admit, however, that relative to the lower mantle, there is a slight westward component in the motions of all the plates. They demonstrated though that the westward drift, seen only for the past 30 Ma, is attributed to the increased dominance of the steadily growing and accelerating Pacific plate. The debate is still open.

Relative significance of each driving force mechanism

The actual vector of a plate's motion must necessarily be a function of all the forces acting upon the plate. However, therein remains the problem regarding what degree each process contributes to the motion of each tectonic plate.

The diversity of geodynamic settings and properties of each plate must clearly result in differences in the degree to which such processes are actively driving the plates. One method of dealing with this problem is to consider the relative rate at which each plate is moving and to consider the available evidence of each driving force upon the plate as far as possible.

One of the most significant correlations found is that lithospheric plates attached to downgoing (subducting) plates move much faster than plates not attached to subducting plates. The Pacific plate, for instance, is essentially surrounded by zones of subduction (the so-called Ring of Fire) and moves much faster than the plates of the Atlantic basin, which are attached (perhaps one could say 'welded') to adjacent continents instead of subducting plates. It is thus thought that forces associated with the downgoing plate (slab pull and slab suction) are the driving forces which determine the motion of plates, except for those plates which are not being subducted (Conrad & Lithgow-Bertelloni 2002). The driving forces of plate motion continue to be active subjects of on-going research within geophysics and tectonophysics.

Historical context - development of the theory

Plate tectonics is the main current theory in Earth Sciences regarding the development of our planet Earth. It is, therefore, appropriate to dedicate some space to explain how the Earth Science community, step by step, has built this theory, from early speculations, through the gathering of proof and severe debates, up to the refinement and quantification, and still ongoing confrontations with alternative ideas.

Summary

Detailed map showing the tectonic plates with their movement vectors

In line with other previous and contemporaneous proposals, in 1912 the meteorologist Alfred Wegener amply described what he called continental drift, expanded in his 1915 book *The Origin of Continents and Oceans* and the scientific debate started that would end up fifty years later in the theory of plate tectonics (Hughes 2001a). Starting from the idea (also expressed by his forerunners) that the present continents once formed a single land mass (which was called Pangea later on) that drifted apart, thus releasing the continents from the Earth's mantle and likening them to "icebergs" of low density granite floating on a sea of denser basalt (Wegener 1966; Hughes 2001b).

But without detailed evidence and a force sufficient to drive the movement, the theory was not generally accepted: the Earth might have a solid crust and mantle and a liquid core, but there seemed to be no way that portions of the crust could move around.

Notwithstanding much opposition, the view of continental drift gained support and a lively debate started between "drifters" or "mobilists" (proponents of the theory) and "fixists" (opponents). During the 1920s, 1930s and 1940s, the former reached important milestones proposing that convection currents might have driven the plate movements, and that spreading may have occurred below the sea within the oceanic crust. Concepts close to the elements now incorporated in plate tectonics were proposed by geophysisists and geologists (both fixists and mobilists) like Vening-Meinesz, Holmes, and Umbgrove.

One of the first pieces of geophysical evidence that was used to support the movement of lithospheric plates came from paleomagnetism. This is based on the fact that rocks of different ages show a variable magnetic field direction, evidenced by studies since the mid–nineteenth century. The magnetic north and south reverse through time, and,

especially important in paleotectonic studies, the relative position of the magnetic north varies through time. Initially, during the first half of the twentieth century, the latter phenomena was explained by introducing what was called "polar wander", i.e., it was assumed that the north pole location had been shifting through time. An alternative explanation, though, was that the continents had moved (shifted and rotated) relative to the north pole, and each continent, in fact, shows its own "polar wander path". During the late 1950s in was shown with success that these data could show the validity of continental drift in two occasions: by Keith Runcorn in a paper in 1956, and by Warren Carey in a symposium held in March 1956.

The second piece of evidence in support of continental drift came during the late 1950s and early 60s from data on the bathymetry of the deep ocean floors and the nature of the oceanic crust such as magnetic properties and, more generally, with the development of marine geology which gave evidence for the association of seafloor spreading along the mid-oceanic ridges and magnetic field reversals, published between 1959 and 1963 by Heezen, Dietz, Hess, Mason, Vine & Matthews, and Morley (Korgen 1995; Spiess & Kuperman 2003).

Simultaneous advances in early seismic imaging techniques in and around Wadati-Benioff zones along the trenches bounding many continental margins, together with many other geophysical (e.g. gravimetric) and geological observations, showed how the oceanic crust could disappear into the mantle, providing the mechanism to balance the extension of the ocean basins with shortening along its margins.

All these evidences, both from the ocean floor and from the continental margins made clear around 1965 that continental drift was feasible and the theory of plate tectonics, which was defined in a series of papers between 1965 and 1967, was born, with all its extraordinary explanatory and predictive power. The theory revolutionized the Earth sciences, explaining a diverse range of geological phenomena and their implications in other studies such as paleogeography and paleobiology.

Continental drift

In the late 19th and early 20th centuries, geologists assumed that the Earth's major features were fixed, and that most geologic features such as basin development and mountain ranges could be explained by vertical crustal movement, described in what is called the geosynclinal theory. Generally, this was placed in the context of a contracting planet Earth due to heat loss in the course of a relatively short geological time.

It was observed as early as 1596 that the opposite coasts of the Atlantic Ocean—or, more precisely, the edges of the continental shelves—have similar shapes and seem to have once fitted together (Kious & Tilling 1996).

Since that time many theories were proposed to explain this apparent complementarity, but the assumption of a solid Earth made these various proposals difficult to accept (Frankel 1987).

The discovery of radioactivity and its associated heating properties in 1895 prompted a re-examination of the apparent age of the Earth (Joly 1909) since this had previously been estimated by its cooling rate and assumption the Earth's surface radiated like a black body (Thomson 1863).

Those calculations had implied that, even if it started at red heat, the Earth would have dropped to its present temperature in a few tens of millions of years. Armed with the knowledge of a new heat source, scientists realized that the Earth would be much older, and that its core was still sufficiently hot to be liquid.

By 1915, after having published a first article in 1912 (Wegener 1912). Alfred Wegener was making serious arguments for the idea of continental drift in the first edition of *The Origin of Continents and Oceans*. In that book (re-issued in four successive editions up to the final one in 1936), he noted how the east coast of South America and the west coast of Africa looked as if they were once attached. Wegener wasn't the first to note this (Abraham Ortelius, Snider-Pellegrini, Roberto Mantovani and Frank Bursley Taylor preceded him just to mention a few), but he was the first to marshal significant fossil and paleo-topographical and climatological evidence to support this simple observation (and was supported in this by researchers such as Alex du Toit). Furthermore, when the rock strata of the margins of separate continents are very similar it suggests that these rocks were formed in the same way, implying that they were joined initially. For instance, some parts of Scotland and Ireland contain rocks very similar to those found in Newfoundland and New Brunswick. Furthermore, the Caledonian Mountains of Europe and parts of the Appalachian Mountains of North America are very similar in structure and lithology.

However, his ideas were not taken seriously by many geologists, who pointed out that there was no apparent mechanism for continental drift. Specifically, they did not see how continental rock could plow through the much denser rock that makes up oceanic crust. Wegener could not explain the force that drove continental drift, and his vindication did not come until after his death in 1930.

Floating continents - paleomagnetism - seismicity zones

As it was observed early that although granite existed on continents, seafloor seemed to be composed of denser basalt, the prevailing concept during the first half of the twentieth century was that there were two types of crust, named "sial" (continental type crust), and "sima" (oceanic type crust). Furthermore, it was supposed that a static shells of strata was present under the continents. It therefore looked apparent that a layer of basalt (sial) underlies the continental rocks.

However, based upon abnormalities in plumb line deflection by the Andes in Peru, Pierre Bouguer had deduced that less-dense mountains must have a downward projection into the denser layer underneath. The concept that mountains had "roots" was confirmed by George B. Airy a hundred years later during study of Himalayan gravitation, and seismic studies detected corresponding density variations. Therefore, by the mid–1950s the

question remained unresolved of whether mountain roots were clenched in surrounding basalt or were floating upon it like an iceberg.

During the 20th century, improvements in and greater use of seismic instruments such as seismographs enabled scientists to learn that earthquakes tend to be concentrated in specific areas, most notably along the oceanic trenches and spreading ridges. By the late 1920s, seismologists were beginning to identify several prominent earthquake zones parallel to the trenches that typically were inclined 40–60° from the horizontal and extended several hundred kilometers into the Earth. These zones later became known as Wadati-Benioff zones, or simply Benioff zones, in honor of the seismologists who first recognized them, Kiyoo Wadati of Japan and Hugo Benioff of the United States. The study of global seismicity greatly advanced in the 1960s with the establishment of the Worldwide Standardized Seismograph Network (WWSSN) to monitor the compliance of the 1963 treaty banning above-ground testing of nuclear weapons. The much improved data from the WWSSN instruments allowed seismologists to map precisely the zones of earthquake concentration world wide.

Meanwhile, debates developed around the phenomena of polar Wander. Since the early debates of continental drift, scientists had discussed and used evidence that polar drift had occurred due to the fact that continents seemed to have moved through different climatic zones during the past. Furthermore, paleomagnetic data had shown that the magnetic pole had also shifted during time. Reasoning in an opposite way, the continents might have shifted and rotated, while the pole remained relatively fixed. The first time the evidence of magnetic polar wander was used to support the movements of continents was in a paper by Keith Runcorn in 1956, and successive papers by him and his students Ted Irving (who was actually the first to be convinced of the fact that paleomagnetism supported continental drift) and Ken Creer.

This was immediately followed by a symposium in Tasmania in March 1956 (Carey 1956;). In this symposium, the evidence was used in the theory of an expansion of the global crust. In this hypothesis the shifting of the continents can be simply explained by a large increase in size of the Earth since its formation. However, this was unsatisfactory because its supporters could offer no convincing mechanism to produce a significant expansion of the Earth. Certainly there is no evidence that the moon has expanded in the past 3 billion years; other work would soon show that the evidence was equally in support of continental drift on a globe with a stable radius.

During the thirties up to the late fifties, numerous milestones were reached that would eventually lead to the development of plate tectonics. These are the works of Vening-Meinesz, Holmes, Umbgrove, and numerous others, in which concepts close or near identical to modern plate tectonics theory where defined and outlined. The most important milestone was reached when the English geologist Arthur Holmes proposed in 1920 that plate junctions might lie beneath the sea, and in 1928 that convection currents within the mantle might be the driving force.

Often, all these milestones are forgotten for various reasons:

1. During this timespan, continental drift was not accepted.
2. Some of these ideas were discussed in the context of abandoned fixistic ideas of a deforming globe without continental drift or an expanding Earth.
3. They were published during an episode of extreme political and economic instability and scientific communication was obviously hampered by this.
4. Many of these were published by European scientists and at first not mentioned or given little credit in the papers published by the American researchers which during the 1960s presented evidence for sea floor spreading.

Mid oceanic ridge spreading and convection

In 1947, a team of scientists led by Maurice Ewing utilizing the Woods Hole Oceanographic Institution's research vessel *Atlantis* and an array of instruments, confirmed the existence of a rise in the central Atlantic Ocean, and found that the floor of the seabed beneath the layer of sediments consisted of basalt, not the granite which is the main constituent of continents. They also found that the oceanic crust was much thinner than continental crust. All these new findings raised important and intriguing questions (Lippsett 2001; and Lippsett 2006).

The new data that had been collected on the ocean basins also showed particular characteristics regarding the bathymetry. One of the major outcomes of these datasets was that all along the globe, a system of mid-oceanic ridges was detected. An important conclusion was that along this system, new ocean floor was being created, which led to the concept of the "Great Global Rift". This was described in the crucial paper of Bruce Heezen (1960) which would trigger a real revolution in thinking. A profound consequence of seafloor spreading is that new crust was, and is now, being continually created along the oceanic ridges. Therefore, Heezen advocated the so-called "expanding Earth" hypothesis of S. Warren Carey (see above). So, still the question remained: how can new crust be continuously added along the oceanic ridges without increasing the size of the Earth? In reality, this question had been solved already by numerous scientists during the forties and the fifties, like Arthur Holmes, Vening-Meinesz, Coates and many others: The crust in excess disappeared along what were called the oceanic trenches where so-called "subduction" occurred. Therefore, when various scientists during the early sixties started to reason on the data at their disposal regarding ocean floor, the pieces of the theory fell quickly at its place.

The question particularly intrigued Harry Hammond Hess, a Princeton University geologist and a Naval Reserve Rear Admiral, and Robert S. Dietz, a scientist with the U.S. Coast and Geodetic Survey who first coined the term *seafloor spreading*. Dietz and Hess (the former published the same idea one year earlier in *Nature*, but priority belongs to Hess who had already distributed an unpublished manuscript of his 1962 article by 1960) were among the small handful who really understood the broad implications of sea floor spreading and how it would eventually agree with the, at that time, unconventional and unaccepted ideas of continental drift and the elegant and mobilistic models proposed by previous workers like Holmes.

In the same year, Robert R. Coats of the U.S. Geological Survey described the main features of island arc subduction in the Aleutian Islands. His paper, though little–noted (and even ridiculed) at the time, has since been called "seminal" and "prescient". In reality, it actually shows that the work by the European scientists on island arcs and mountain belts performed and published during the 1930s up until the 1950s was applied and appreciated also in the United States.

If the Earth's crust was expanding along the oceanic ridges, Hess and Dietz reasoned like Holmes and others before them, it must be shrinking elsewhere. Hess followed Heezen suggesting that new oceanic crust continuously spreads away from the ridges in a conveyor belt–like motion. And, using the mobilistic concepts developed before, he correctly concluded that many millions of years later, the oceanic crust eventually descends along the continental margins where oceanic trenches – very deep, narrow canyons are present e.g. along the rim of the Pacific Ocean basin – were formed. The important step Hess made was that convection currents would be the driving force in this process, arriving at the same conclusions as Holmes had decades before with the only difference that the thinning of the ocean crust was performed using the mechanism of Heezen of spreading along the ridges. Hess therefore concluded that the Atlantic Ocean was expanding while the Pacific Ocean was shrinking. As old oceanic crust is "consumed" in the trenches, (like Holmes and others, he believed this was done by thickening of the continental lithosphere, not, as nowadays believed, by underthrusting at a larger scale of the oceanic crust itself into the mantle) new magma rises and erupts along the spreading ridges to form new crust. In effect, the ocean basins are perpetually being "recycled," with the creation of new crust and the destruction of old oceanic lithosphere occurring simultaneously, in a way like what later would be called the Wilson cycle (see below). Thus, the new mobilistic concepts neatly explained why the Earth does not get bigger with sea floor spreading, why there is so little sediment accumulation on the ocean floor, and why oceanic rocks are much younger than continental rocks.

The final proof: magnetic striping
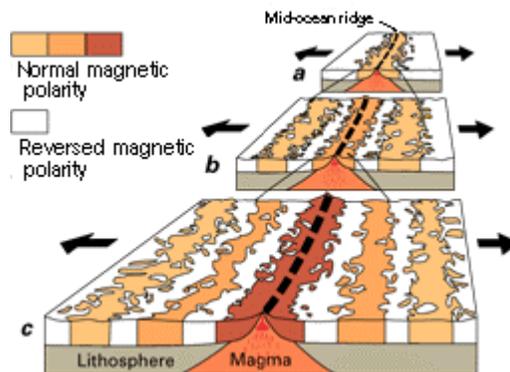
Seafloor magnetic striping
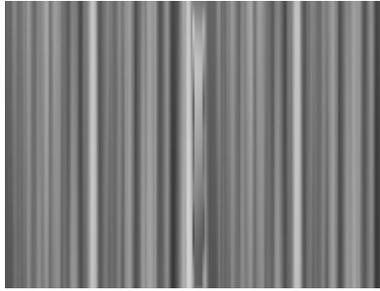
A demonstration of magnetic striping. (The darker the color is the closer it is to normal polarity)

Beginning in the 1950s, scientists like Victor Vacquier, using magnetic instruments (magnetometers) adapted from airborne devices developed during World War II to detect submarines, began recognizing odd magnetic variations across the ocean floor. This finding, though unexpected, was not entirely surprising because it was known that basalt—the iron-rich, volcanic rock making up the ocean floor—contains a strongly magnetic mineral (magnetite) and can locally distort compass readings. This distortion was recognized by Icelandic mariners as early as the late 18th century. More important, because the presence of magnetite gives the basalt measurable magnetic properties, these newly discovered magnetic variations provided another means to study the deep ocean floor. When newly formed rock cools, such magnetic materials recorded the Earth's magnetic field at the time.

As more and more of the seafloor was mapped during the 1950s, the magnetic variations turned out not to be random or isolated occurrences, but instead revealed recognizable patterns. When these magnetic patterns were mapped over a wide region, the ocean floor showed a zebra-like pattern: one stripe with normal polarity and the adjoining stripe with reversed polarity. The overall pattern, defined by these alternating bands of normally and reversely polarized rock, became known as magnetic striping, and was published by Ron G. Mason and co-workers in 1961, who didn't find, though, an explanation for these data in terms of sea floor spreading, like Vine, Matthews and Morley a few years later (Mason & Raff 1961); (Raff & Mason 1961).

The discovery of magnetic striping called for an explanation. In the early 1960s scientists such as Heezen, Hess and Dietz had begun to theorise that mid-ocean ridges mark structurally weak zones where the ocean floor was being ripped in two lengthwise along the ridge crest. New magma from deep within the Earth rises easily through these weak zones and eventually erupts along the crest of the ridges to create new oceanic crust. This process, at first denominated the "conveyer belt hypothesis" and later called seafloor spreading, operating over many millions of years continues to form new ocean floor all across the 50,000 km-long system of mid–ocean ridges.

Only four years after the maps with the "zebra pattern" of magnetic stripes were published, the link between sea floor spreading and these patterns was correctly placed, independently by Lawrence Morley, and by Fred Vine and Drummond Matthews, in 1963 (Vine & Matthews 1963) now called the Vine-Matthews-Morley hypothesis. This

hypothesis linked these patterns to geomagnetic reversals and was supported by several lines of evidence:

1. the stripes are symmetrical around the crests of the mid-ocean ridges; at or near the crest of the ridge, the rocks are very young, and they become progressively older away from the ridge crest;
2. the youngest rocks at the ridge crest always have present-day (normal) polarity;
3. stripes of rock parallel to the ridge crest alternate in magnetic polarity (normal-reversed-normal, etc.), suggesting that they were formed during different epochs documenting the (already known from independent studies) normal and reversal episodes of the Earth's magnetic field.

By explaining both the zebra-like magnetic striping and the construction of the mid-ocean ridge system, the seafloor spreading hypothesis (SFS) quickly gained converts and represented another major advance in the development of the plate-tectonics theory. Furthermore, the oceanic crust now came to be appreciated as a natural "tape recording" of the history of the geomagnetic field reversals (GMFR) of the Earth's magnetic field. Nowadays, extensive studies are dedicated to the calibration of the normal-reversal patterns in the oceanic crust on one hand and known timescales derived from the dating of basalt layers in sedimentary sequences (magnetostratigraphy) on the other, to arrive at estimates of past spreading rates and plate reconstructions.

Definition and refining of the theory - from new global tectonics to plate tectonics

After all these considerations, Plate Tectonics (or, as it was initially called "New Global Tectonics") became quickly accepted in the scientific world, and numerous papers followed that defined the concepts:

- In 1965, Tuzo Wilson who had been a promotor of the sea floor spreading hypothesis and continental drift from the very beginning (e.g. Wilson 1963) added the concept of transform faults to the model, completing the classes of fault types necessary to make the mobility of the plates on the globe work out (Wilson 1965).
- A symposium on continental drift was held at the Royal Society of London in 1965 which must be regarded as the official start of the acceptance of plate tectonics by the scientific community, and which abstracts are issued as Blacket, Bullard & Runcorn (1965). In this symposium, Edward Bullard and co-workers showed with a computer calculation how the continents along both sides of the Atlantic would best fit to close the ocean, which became known as the famous "Bullard's Fit".
- In 1966 Tuzo Wilson published the paper that referred to previous plate tectonic reconstructions, introducing the concept of what is now known as the "Wilson Cycle" (Wilson 1966).
- In 1967, at the American Geophysical Union's meeting, W. Jason Morgan proposed that the Earth's surface consists of 12 rigid plates that move relative to each other (Morgan 1968).

- Two months later, Xavier Le Pichon published a complete model based on 6 major plates with their relative motions, and we may say that this parks the final acceptance of the scientific community of plate tectonics (Le Pichon 1967).
- In the same year, McKenzie and Parker independently presented a model similar to Morgan's using translations and rotations on a sphere to define the plate motions (McKenzie & Parker 1967).

Implications for biogeography

Continental drift theory helps biogeographers to explain the disjunct biogeographic distribution of present day life found on different continents but having similar ancestors (Moss & Wilson 1998). In particular, it explains the Gondwanan distribution of ratites and the Antarctic flora.

Plate reconstruction
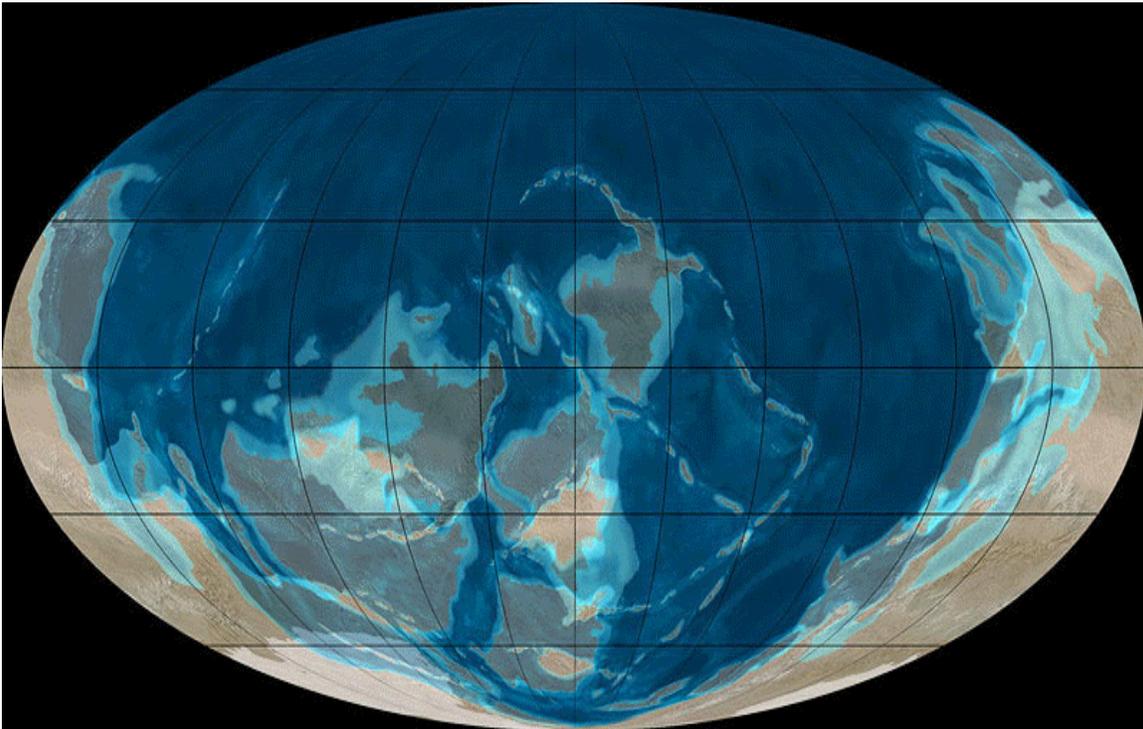
Reconstruction of plate configurations for the whole Phanerozoic

Reconstruction is used to establish past (and future) plate configurations, helping determine the shape and make-up of ancient supercontinents and providing a basis for paleogeography.
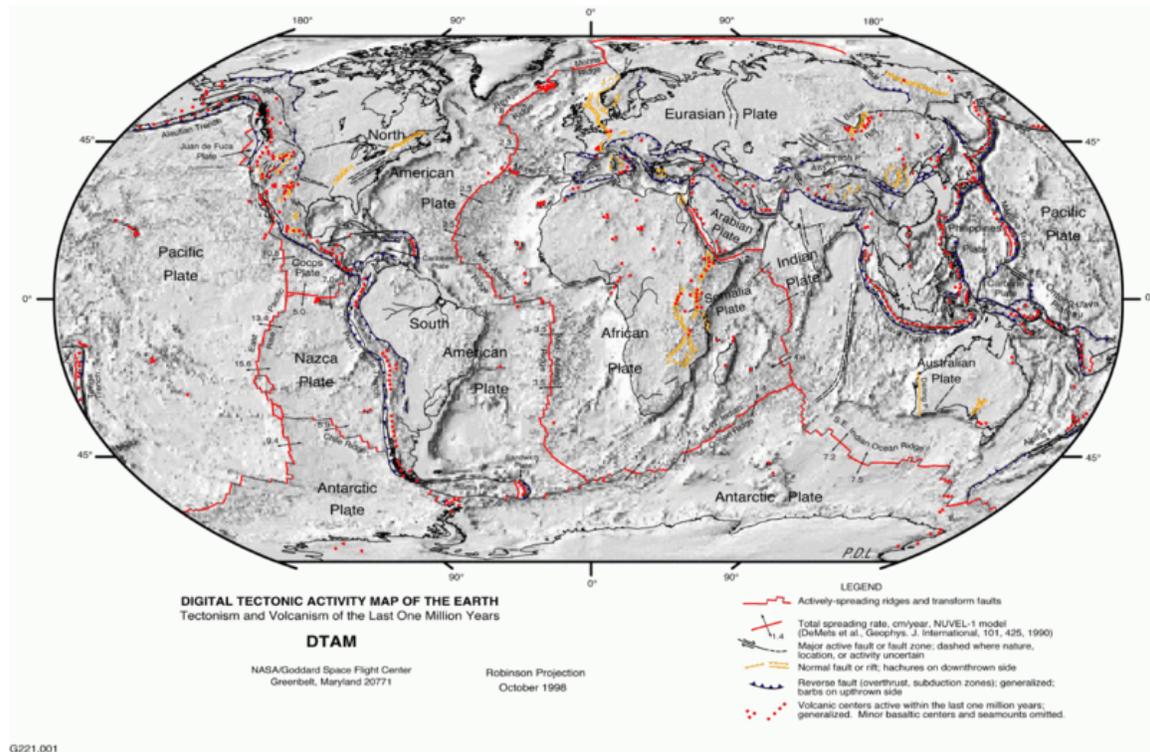
Defining plate boundaries

Current plate boundaries are defined by their seismicity (Condie 1997). Past plate boundaries within existing plates are identified from evidence of vanished oceans, such as ophiolites (Lliboutry 2000).

Past plate motions

The movement of plates has caused the formation and break-up of continents over time, including occasional formation of a supercontinent that contains most or all of the continents. The supercontinent Rodinia is thought to have formed about 1 billion years ago and to have embodied most or all of Earth's continents, and broken up into eight continents around 600 million years ago. The eight continents later re-assembled into another supercontinent called Pangaea; Pangaea broke up into Laurasia (which became North America and Eurasia) and Gondwana (which became the remaining continents).

Various types of quantitative and semi-quantitative information are available to constrain past plate motions. The geometric fit between continents, such as between west Africa and South America is still an important part of plate reconstruction. Magnetic stripe patterns provide a reliable guide to relative plate motions going back into the Jurassic period. The tracks of hotspots give absolute reconstructions but these are only available back to the Cretaceous (Torsvik 2008). Older reconstructions rely mainly on paleomagnetic pole data, although these only constrain the latitude and rotation, but not the longitude. Combining poles of different ages in a particular plate to produce apparent polar wander paths provides a method for comparing the motions of different plates through time (Butler 1992). Additional evidence comes from the distribution of certain sedimentary rock types, faunal provinces shown by particular fossil groups, and the position of orogenic belts (Torsvik 2008).

Current plates

DIGITAL TECTONIC ACTIVITY MAP OF THE EARTH
Tectonism and Volcanism of the Last One Million Years

DTAM

NASA/Goddard Space Flight Center
Greenbelt, Maryland 20771

Robinson Projection
October 1998

Major plates

Depending on how they are defined, there are usually seven or eight "major" plates:

- African Plate
- Antarctic Plate
- Indo-Australian Plate, sometimes subdivided into:
  - Indian Plate
  - Australian Plate
- Eurasian Plate
- North American Plate
- South American Plate
- Pacific Plate

Minor plates

There are dozens of smaller plates, the seven largest of which are:

- Arabian Plate
- Caribbean Plate
- Juan de Fuca Plate
- Cocos Plate

- Nazca Plate
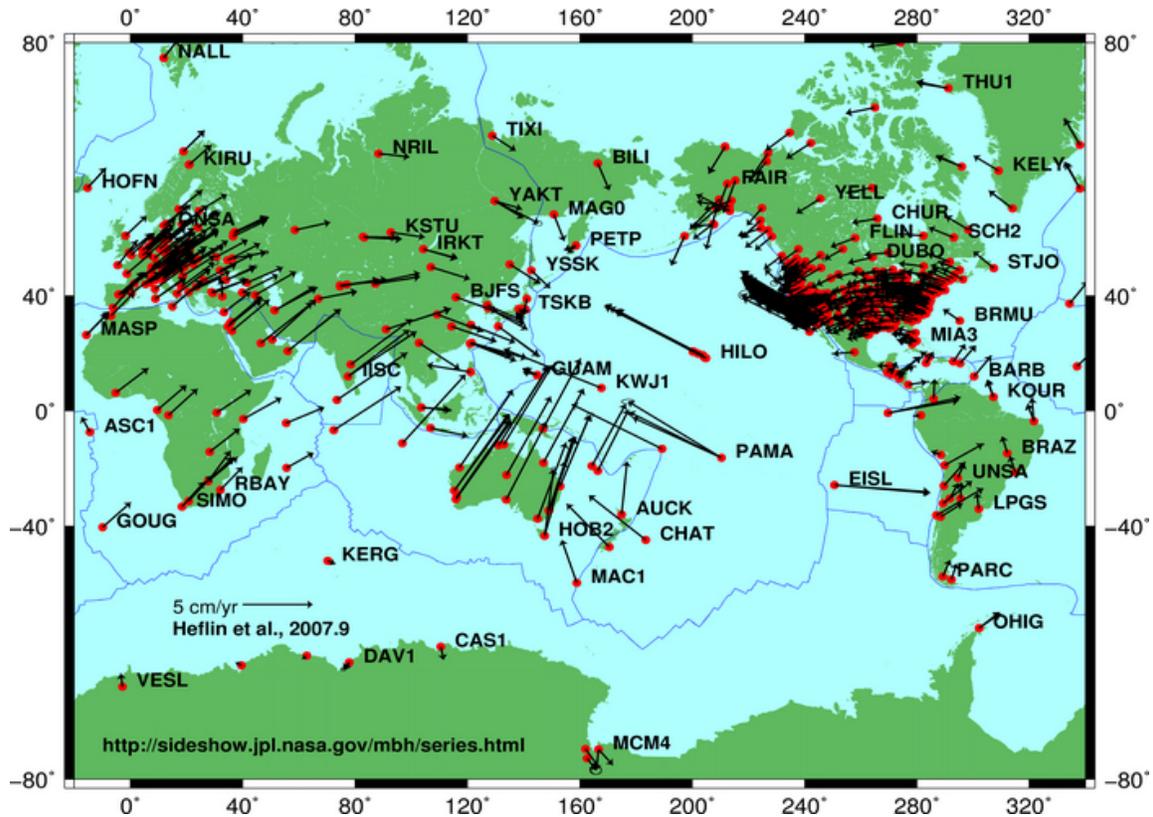- Philippine Sea Plate
- Scotia Plate

Current Motion

Plate motion based on Global Positioning System (GPS) satellite data from NASA JPL. The vectors show direction and magnitude of motion.

The current motion of the tectonic plates is nowadays revealed from remote sensing satellite data sets, calibrated with ground station measurements.

Plate tectonics on other celestial bodies (Planets, Moons)

The appearance of plate tectonics on terrestrial planets is related to planetary mass, with more massive planets than Earth expected to exhibit plate tectonics. Earth may be a borderline case, owing its tectonic activity to abundant water (Valencia, O'Connell & Sasselov 2007)

(Silica and water form a deep eutectic.)

Venus

Venus shows no evidence of active plate tectonics. There is debatable evidence of active tectonics in the planet's distant past; however, events taking place since then (such as the plausible and generally accepted hypothesis that the Venusian lithosphere has thickened greatly over the course of several hundred million years) has made constraining the course of its geologic record difficult. However, the numerous well-preserved impact craters have been utilized as a dating method to approximately date the Venusian surface (since there are thus far no known samples of Venusian rock to be dated by more reliable methods). Dates derived are dominantly in the range c. 500 to 750 Ma, although ages of up to c. 1.2 Ga have been calculated. This research has led to the fairly well accepted hypothesis that Venus has undergone an essentially complete volcanic resurfacing at least once in its distant past, with the last event taking place approximately within the range of estimated surface ages. While the mechanism of such an impressive thermal event remains a debated issue in Venusian geosciences, some scientists are advocates of processes involving plate motion to some extent.

One explanation for Venus' lack of plate tectonics is that on Venus temperatures are too high for significant water to be present (Kasting 1988). The Earth's crust is soaked with water, and water plays an important role in the development of shear zones. Plate tectonics requires weak surfaces in the crust along which crustal slices can move, and it may well be that such weakening never took place on Venus because of the absence of water. However, some researchers remain convinced that plate tectonics is or was once active on this planet.

Mars

Mars is considerably smaller than Earth and Venus, and there is evidence for ice on its surface and in its crust.

In the 1990s, it was proposed that Martian Crustal Dichotomy was created by plate tectonic processes (Sleep 1994). Scientists today disagree, and believe that it was created either by upwelling within the Martian mantle that thickened the crust of the Southern Highlands and formed Tharsis (Zhong & Zuber 2001) or by a giant impact that excavated the Northern Lowlands (Andrews-Hanna, Zuber & Banerdt 2008).

Observations made of the magnetic field of Mars by the *Mars Global Surveyor* spacecraft in 1999 showed patterns of magnetic striping discovered on this planet. Some scientists interpreted these as requiring plate tectonic processes, such as seafloor spreading (Connerney et al. 1999, Connerney et al. 2005)). However, their data fail a "magnetic reversal test", which is used to see if they were formed by flipping polarities of a global magnetic field (Harrison 2000).

Galilean satellites of Jupiter

Some of the satellites of Jupiter have features that may be related to plate-tectonic style deformation, although the materials and specific mechanisms may be different from plate-tectonic activity on Earth.

Titan, moon of Saturn

Titan, the largest moon of Saturn, was reported to show tectonic activity in images taken by the Huygens Probe, which landed on Titan on January 14, 2005 (Soderblom et al. 2007).