ETH Zürich - Geophysics

Plate Tectonics

The tectonics of lithospheric plates

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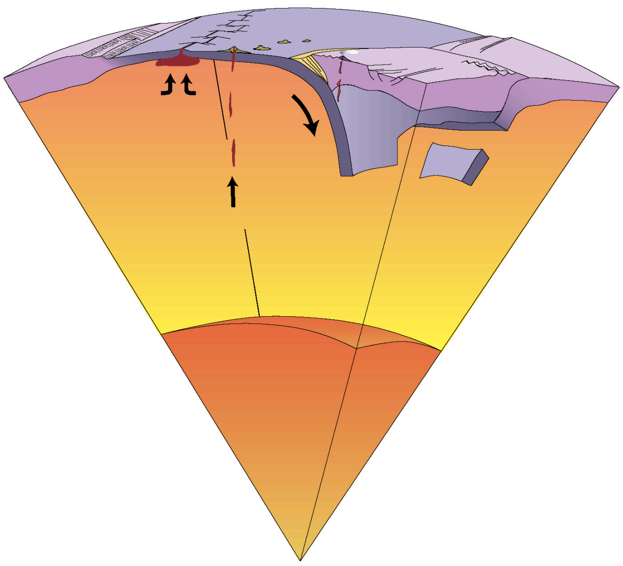

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1 The thermal structure of the Earth

Plate tectonics is an expression of the Earth's long term loss of internal heat. In a first approximation the Earth comprises two different liquids, the core with a diameter of ca. 3200 km and the mantle with a thickness of ca. 2900 km which surrounds the core. Since the temperature outside the Earth is far below the melting temperature of the mantle material (about 1250°C for peridotitic rock), the outermost layer of the mantle is frozen. We call this solid rock layer having a thickness of ca. 100 km the lithosphere. The lithosphere surrounds the viscous mantle and thermally isolates it completely from the Earth's cold upper atmosphere (Fig. 1.1).

If the inner earth did not continue to produce heat, the lithosphere would slowly increase in thickness until finally the whole mantle would be frozen. The slow but steady decay of the natural radioactive elements uranium and thorium in the Earth acts as an internal long-term heat source. A second heat source for the Earth's mantle consists in the slow freezing of the core. Under the prevailing extreme high pressures, the melting temperature of the core material (a mixture of iron-nickel-sulfur-oxygen) is well above 5000°C. The inner core is solid and the liquid outer core freezes against the inside so that the radius of the inner core slowly increases. When the core material freezes, a large amount of energy is set free, which can only escape towards the outside, i.e.

from the core through the mantle to the surface of the earth.

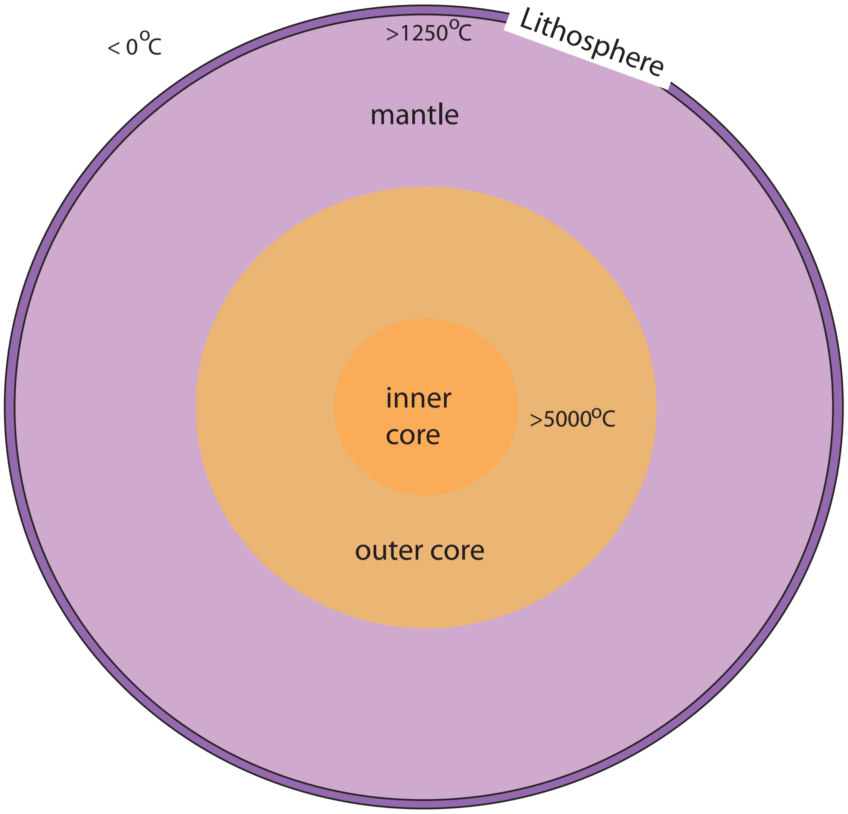


Figure 1.1: The viscous mantle surrounds the outer liquid and solid inner core and is itself surrounded by a thin solid layer, the lithosphere.

Heat transfer from the hot interior of the Earth through the mantle to the Earth's surface is accomplished in different ways, depending on pressure, temperature, the state of matter (solid-liquid-gas), and the type of heat sources (see Appendix 1, heat transport equation). In a solid such as the lithosphere solely heat is transferred almost exclusively by conduction. In contrast, heat transport in fluids occurs primarily by material movement. Hot material rises, and cold material sinks, see Fig. 1.2). This is called convection. Whereas the mantle reacts to short-term loads, such as seismic waves, as a rigid body, it behaves as a liquid with respect to slow movements of a few cm per year. Because of the relatively low efficiency of heat conduction, heat transport in the viscous mantle is dominated by convection. The lithosphere, however, is often referred to as thermal boundary layer (that is, a layer in which heat conduction is the dominant form of heat transport) and isolates the hot mantle (Fig. 1.1 and 1.2).

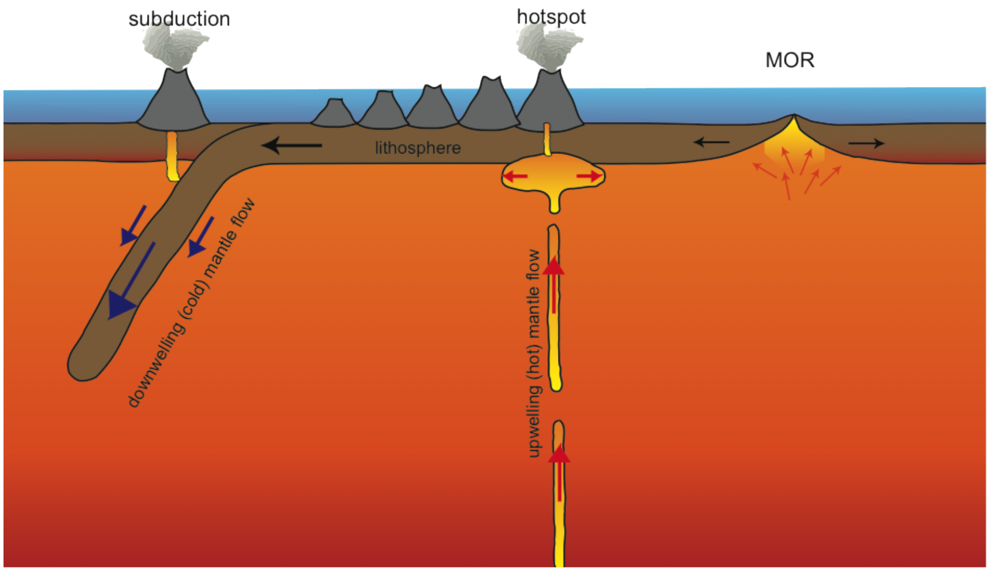


Figure 1.2: The lithosphere-mantle system where cold, dense material (oceanic lithosphere) sinks in subduction zones and hot material in the form of plumes rise up through the mantle reaching the underside of the lithosphere at hot spot locations. The volcanic chain of islands like Hawaii is an imprint on the Earth's surface of such a rising plume (hot spot) below a drifting lithosphere. Oceanic lithosphere is generated on both sides of a mid-ocean ridge (MOR) (see text). Note how rising and sinking mantel flows are only indirectly coupled but do not form a single circular flow pattern.

Convection in the mantle is driven by heat from core, heat production from decay of natural radioactive elements distributed in the mantle, and by the Earth's gravitational field. At constant chemical composition and without phase transitions hot material is less dense than cold material. In a liquid, it needs only small differences in density to trigger equilibrating currents. Locally heated material expands, rises, and induces a counter flow of colder material towards the core.

In a sauce pan the bottom is heated from below while the liquid is cooled from above and the sides. For this reason, normally a circular flow of ascending hot water forms in the middle of the pot and a return flow of cooler water sinks down along the sides. This simple convection model does not, however, apply to the Earth (see chapter on driving forces of plate tectonics later), because the Earth's mantle has no sides rather only a cold upper surface and a hot lower surface hot (core-mantle interface).

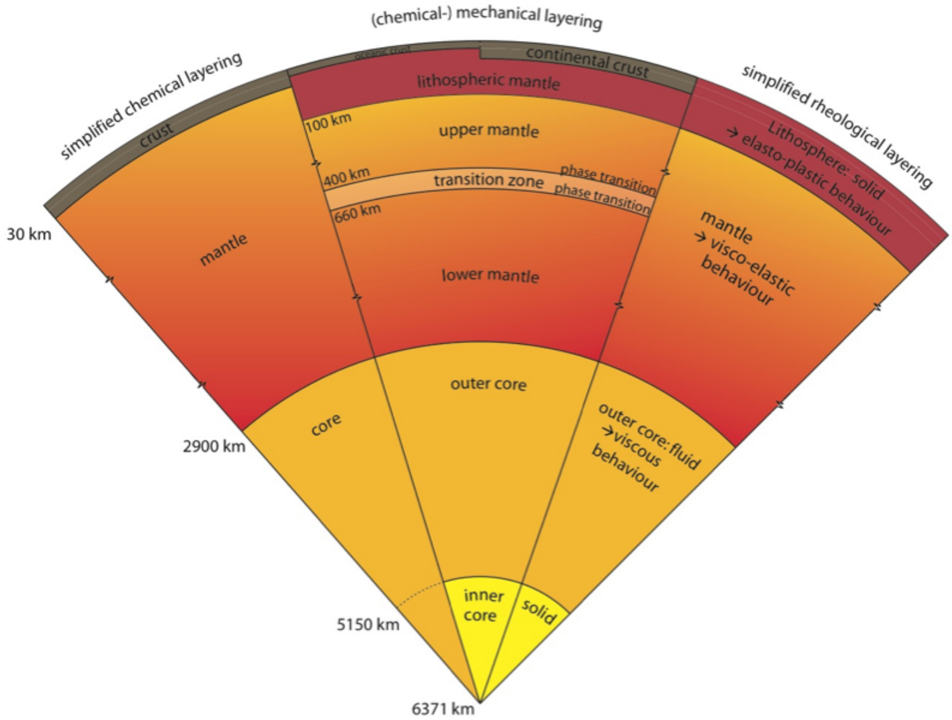


Figure 1.3: The Earth's structure and state. According to chemical composition and density, one differentiates the crust, mantle, and core (left). Seen mechanically the solid lithosphere floats on the viscous mantle that in turn rests on the liquid outer core that surrounds the solid inner core (right). Crucial for understanding the processes of plate tectonics is the combination of the two perspectives (center).

2 The lithosphere floats on the asthenosphere

The solid outer shell, the lithosphere, which floats on the underlying viscous mantle, consists mainly of mantle material with a thin layer of crust material on top (Fig. 1 . 3). Most parts of the lithosphere approximate 100 km in thickness, varying from a minimum of 5 km at mid-ocean ridges (MORs, Fig. 1.2) to a maximum of 250 km (cratons, see below). The lithosphere-asthenosphere system (LAS) of the Earth is analogous to an ice-covered lake in the middle of winter. The ice may vary in thickness, but no water appears found at the surface because it would immediately freeze.

In analogy, the asthenosphere is nowhere exposed on the Earth's surface (Fig. 2.1), even at MORs where the lithosphere is thinnest and comprises only oceanic crust. The reason that oceanic crust is formed at the MOR is due to the property of the asthenosphere, as a complex chemical mixture, at low pressure and the corresponding temperature, to separate into different components. This in a similar way to milk, which can separate into a more watery part and a more creamy part.

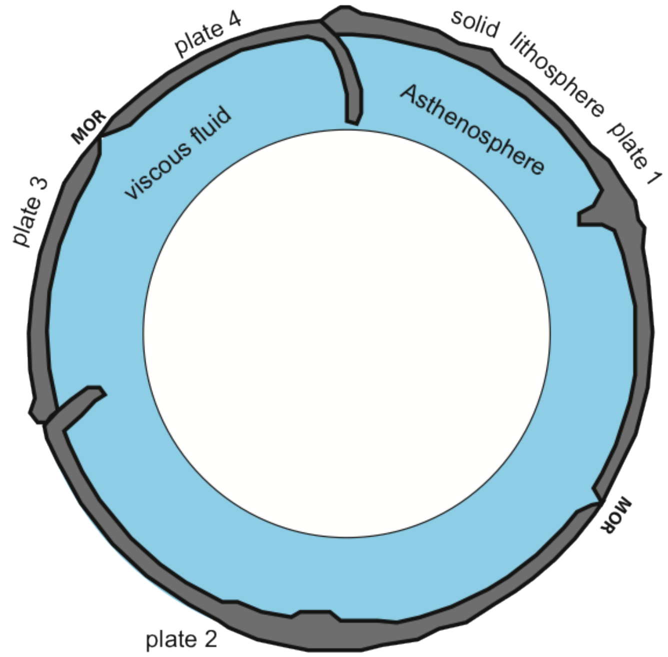


Figure 2.1: The lithosphere is broken into a dozen larger and smaller plates which float on the underlying viscous asthenosphere. The moving lithosphere-plates cover the entire Earth's surface.

In the case of peridotitic[1] asthenosphere arise as a so-called "decompression melting" basaltic magmas. Due to their lower density and much lower viscosity – relative to the surrounding more Fe-rich material – those basaltic magma rise in the asthenosphere under the MOR and freeze as new oceanic floor. This chemical differentiation process (basaltic decompression melts ascending and forming new oceanic crust) are found wherever two lithospheric plates drift away from each other and forming a new MOR, analogous to ocean water upwelling and freezing at the surface between two ice sheets.

In floating solids there exists an isostatic balance between weight of the solid and its buoyancy due to displaced liquid (principle of Archimedes). In a gravity field as such of the earth, the free surface of any liquid automatically corresponds to an equipotential surface. On the earth, the so-called geoid corresponds to the surface of the sea. However, the sea surface does not depict the proper reference level for the isostasy of the LAS, which must refer to the fluid asthenosphere. Whereas you can use the sea surface as a

1. That is comprising mostly the mineral peridotiteFeO_4MgTi.

reference level for the calculation of the topographic load and buoyancy of a single iceberg floating, this procedure cannot be used for a completely frozen lake or even for the asthenosphere completely covered by lithosphere plates. In this case, we must set the reference surface in the liquid (asthenosphere) below the deepest lithosphere and calculate the total mass balance of the overlying material column. This reference surface has the same shape (topography) as the geoid, reflecting how the hypothetical surface of the liquid asthenosphere would look if there were no lithosphere.

The sum of all the masses in each column above the reference surface is the same for all columns. Writing this mass balance for a unit surface:

$$\rho_{\text{topo}} \cdot h_{\text{topo}} + \rho_c \cdot h_c + \rho_A \cdot h_A = \text{const} \tag{2.1}$$

Where ρ_{topo} (= 2.67 g/cm^3) denotes the density and h_{topo} the thickness of the topographic load (mountains), c is the abbreviation for crust, ML the mantle-lithosphere, and A the asthenosphere. See below for the corresponding density values.

The lithosphere is divided up a dozen large and a few small plates that float and move around on the underlying viscous asthenosphere. The solid lithosphere comprises two layers that are bonded to each other: the Earth's crust above and the mantle lithosphere below the Mohorovicic discontinuity, abbreviated Moho (Figure 2.2). Crustal rocks have formed by several processes of chemical differentiation and are characterized by lower density and lower seismic wave velocities in comparison to the underlying peridotitic mantle rocks. The lower lithosphere layer, called the mantle lithosphere, consists of peridotite like the viscous asthenosphere, from which it is formed by freezing (at about 1300°C).

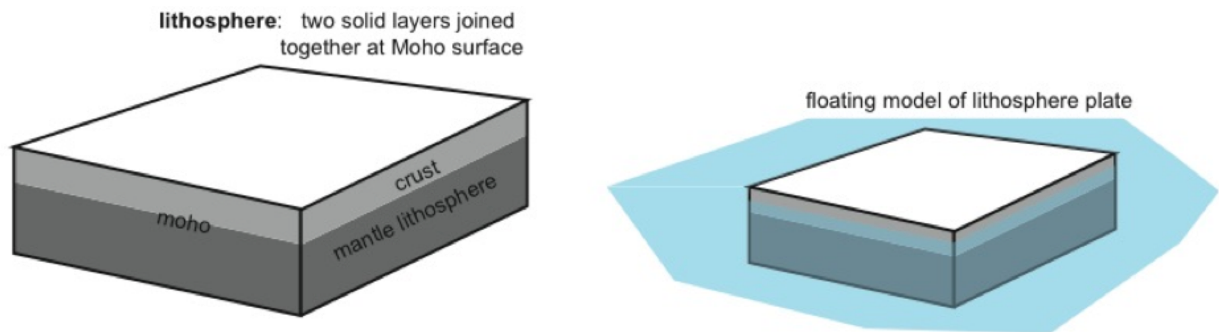


Figure 2.2: The lithosphere comprises two solid layers, the crust and mantle lithosphere that are bonded together along the Moho surface. A lithospheric plate can be considered as raft that floats on the asthenosphere.

Frozen peridotite (mantle lithosphere has a density ρ_{ML} = 3.3 g/cm^3) is denser than the melted peridotite of the asthenosphere (ρ_A = 3.25 g/cm^3). Hence, the solid lithosphere consists of two differently buoyant layers: The upper layer, the crust (continental crust $\rho = 2.85$ g/cm^3, oceanic crust $\rho = 2.9$ g/cm^3), is significantly lighter and the lower layer (ML) is slightly heavier than the underlying viscous liquid. One can, therefore, consider the lithospheric plates as a rafts (Fig. 2.2), consisting of as cork- like upper layer and a metal-like lower layer. As long as the two layers are strongly bound together, the thickness ratio of the lighter upper to the lower heavier layer determines whether the raft floats or sinks.

From the mass balance (2.1) we can derive the conditions for a lithospheric plate with the surface with height [m] h_{topo} (assuming flat topography):

$\rho_{\text{topo}}h_{\text{topo}} + (\rho_c - \rho_A)h_c + (\rho_{ML} - \rho_A)h_{ML} = 0$	plate just balanced	(2.2a)
$\rho_{\text{topo}}h_{\text{topo}} + (\rho_c - \rho_A)h_c + (\rho_{ML} - \rho_A)h_{ML} < 0$	plate floats	(2.2b)
$\rho_{\text{topo}}h_{\text{topo}} + (\rho_c - \rho_A)h_c + (\rho_{ML} - \rho_A)h_{ML} > 0$	plate can sink	(2.2c)

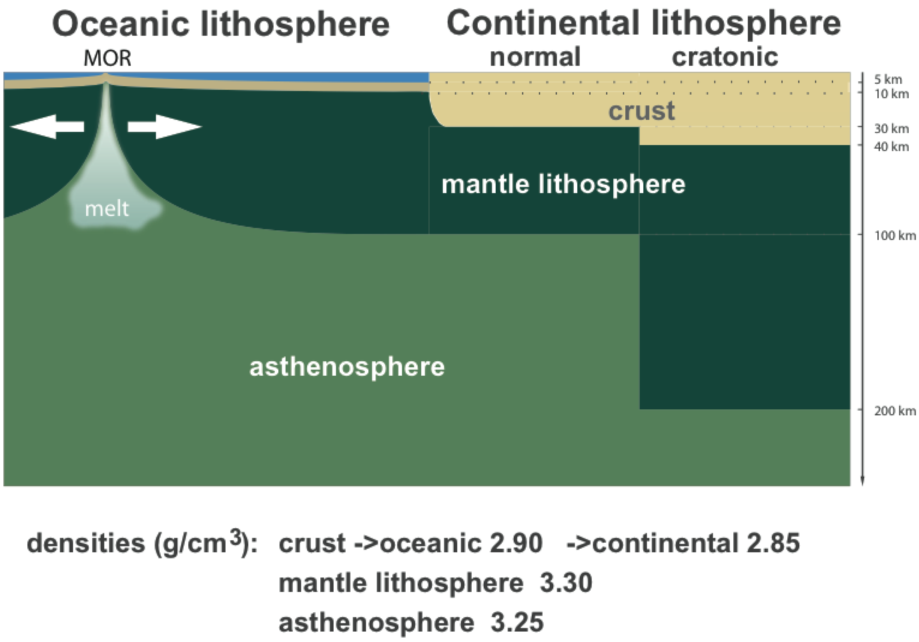


Figure 2.3: Different types of lithosphere and their general structure. Cratonic lithosphere is more than 1700 million years old whereas oceanic and normal continental lithosphere are forming today. A 95 km thick oceanic lithosphere can reach an average density of 3.28 t/m^3 and a 100 km-thick continental lithosphere an average density of 3.16 t/m^3.

Note that on land (continental lithosphere) the density of rock lying above sea level (the topography jutting out above seal level) is assumed to have the density ρ_{topo} = 2.67 g/cm^3 whereas in the case of oceanic lithosphere the topography corresponds to the average water depth h_{topo} and the density of the water column is measured relative to the density of the asthenosphere ($\rho_{\text{topoOcean}}$ = (ρ_{H2O} − ρ_A) = −2.25

g/cm^3).

This means that the oceans with their lower density in the water column relative to crustal rocks are responsible for isostatic equilibrium of the oceanic lithosphere (Fig. 2.3). On the other hand, sea water does not adhere to the underlying lithosphere and thus oceanic lithosphere can sink into the asthenosphere (equation 2.2c) while continental lithosphere floats long time on top because of the 30 km continental crust with an average density less than that of the asthenosphere (equation 2.2b) adhering to mantle lithosphere. The different density ratios relative to the asthenosphere cause the large age-difference of the two lithosphere types: While the oldest oceanic lithosphere is less than 200 million years old, there still exist continental (cratonic) lithosphere plate part that are more than 3,000 million years old.

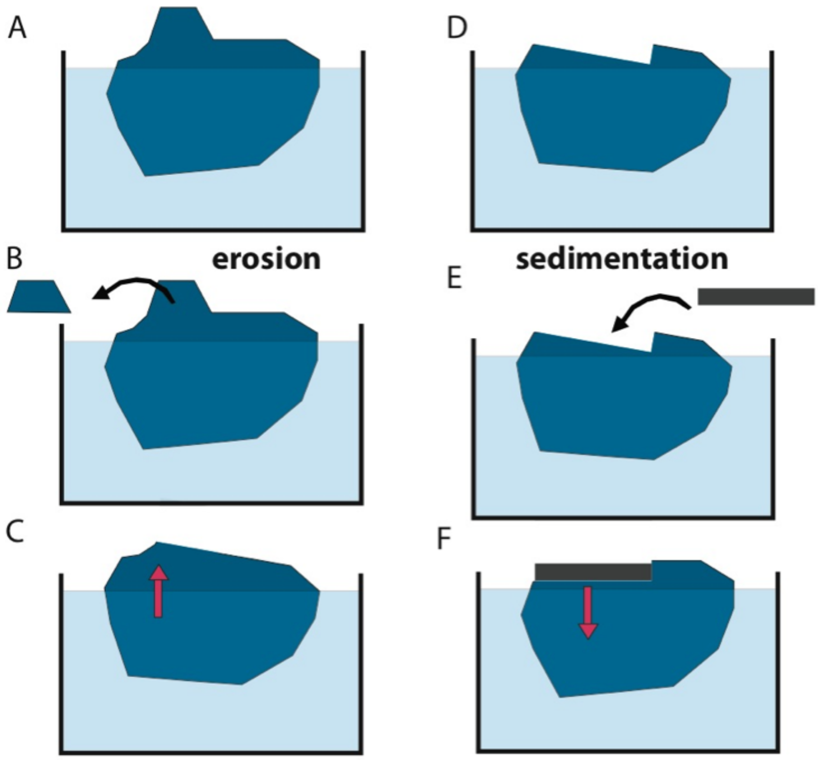


Figure 2.4: The dynamic buoyancy equilibrium of the lithosphere floating on the asthenosphere (often referred to simply as isostasy) is also expressed in local tectonic processes. A mountain range rises when topographic mass is removed by erosion (se- quence A-C) and the deposition of sediments in a basin – as a consequence of erosion in neighboring mountain range – leads to a subsidence (sequence D-F).

The principle of the lithosphere plates acting as rafts floating on the asthenosphere can also help understand two major tectonic processes on continents (Fig. 2.4). Removal of erosion debris by rivers and glaciers decreases the mass of a mountain range and thus reduces the entire static load of the topography on the lithospheric plate. This triggers an isostatic adjustment reflected by again uplifting the mountains while they are eroded. This dynamic can be observed, for example, in the Alps, which are rising at a rate of almost 2 mm / year. Conversely, the filling of a sedimentary basin with detritus from a nearby mountain range leads to a slow but continuous subsidence, which in turn favors more basin infill. As a result, in such basins we often observe a surprising constancy of the grain size of the sedimentary deposits over several kilometers depth. Examples of such sedimentary basins are the Molasse Basin in the north and the Po Basin in the south of the Alps.

Most large lithospheric plates today comprise both continental and oceanic lithosphere which are welded together at a passive continental margin (Fig. 2.5). The entire plate moves on the asthenosphere floating as a unit and responding to changes of vertical loads (isostatic adjustment movements). However, major differences between the oceanic and the continental lithosphere parts a plate can be observed.

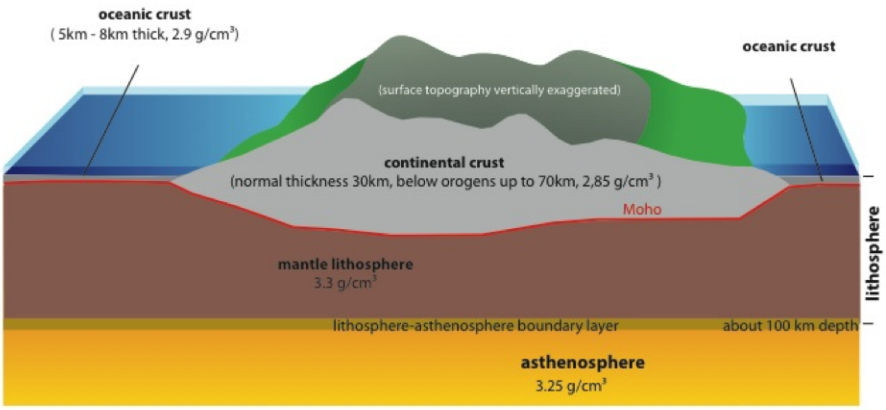


Figure 2.5: Schematic example of a lithospheric plate comprising both continental and oceanic lithosphere (topography is greatly exaggerated). The plate boundaries lie outside the image at two mid-ocean ridges (MORs). Along the both passive continental margins one sees not only the transition from shelf to deep sea regions but also the dramatic reduction in the thickness of the crust from ca. 30 km below the continent with low topography (up to 50 km directly beneath mountains) to a mere 5 km depth (to max. 8 km) beneath the oceans.

Since the lithospheric plates float on the asthenosphere like rafts, the height of the earth's surface gives information about the type of the lithosphere (Fig. 2.6). In addition to the continents the shelf regions also consist of continental lithosphere.

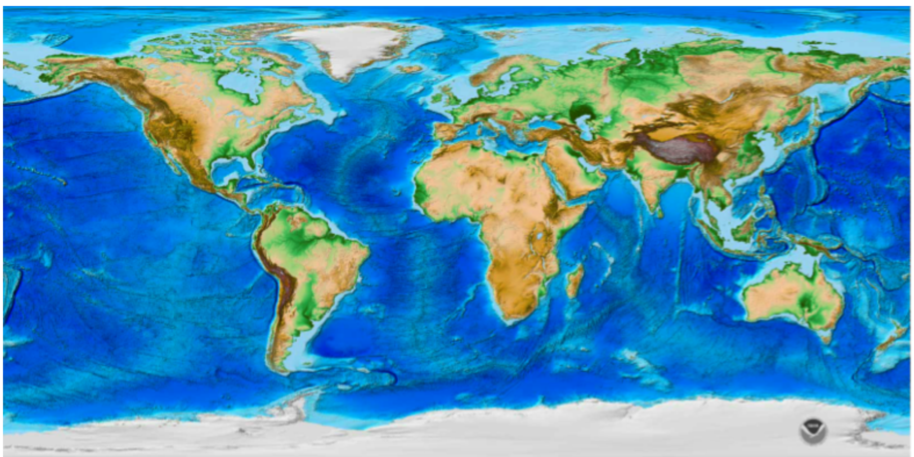


Figure 2.6: Topography of the earth's surface. Bright blue are the shelf areas that consist as the continents themselves from continental lithosphere (National Oceanic and Atmospheric Administration NOAA).

3 The oceanic lithosphere cycle

As a map of the surface heat flux of the earth shows (Fig. 3.1), the earth loses most of its internal heat in the vicinity of the mid-ocean ridges (MORs) where new oceanic lithosphere is created. Elsewhere the lithosphere forms a well insulating layer and a relatively small amount of heat is lost. The large amount of heat lost in the formation of new oceanic lithosphere (especially along and in the vicinity of the MORs, but also when this new lithosphere cools) follows a process analogous to the transition of water to ice, and is accompanied by large release of energy (Fig. 3.2).

Much of the heat from the Earth’s interior is output to the atmosphere at the MOR when fresh asthenospheric material is cooled and forms oceanic lithosphere. This can only continue, however, when the two newly formed oceanic lithosphere plates on either side of the MOR drift apart. Because the size of the Earth remains constant an area equal to the new lithosphere formed must per unit time be removed somewhere else. This occurs where lithosphere is subducted. Since a few million years after its creation the oceanic lithosphere exhibits a density greater than the underlying asthenosphere, and the continental lithosphere always has a lower density than the asthenosphere, oceanic lithosphere is subducted.

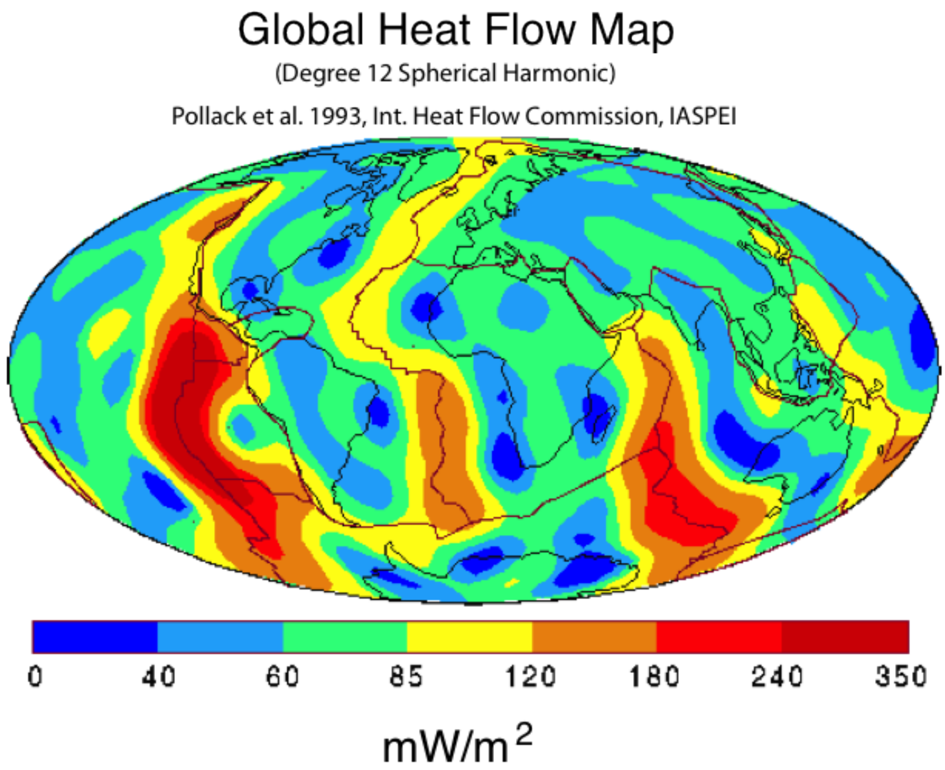


Figure 3.1: Map of the surface heat flux of the Earth compiled by Pollack et al. 1993. Areas with heat flow higher than 100 mW/m^2 are limited to those regions in the immediate vicinity of the mid-ocean ridges.

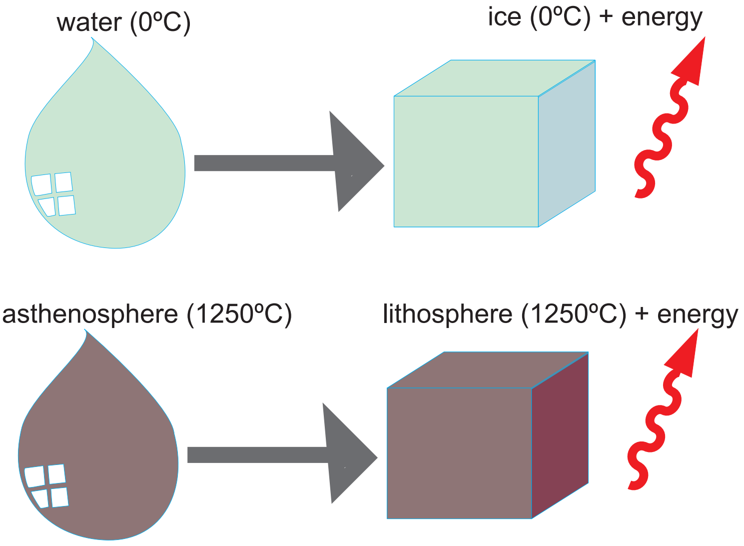


Figure 3.2: In analogy to thee formation of ice from water, the liquid→solid phase transition of the oceanic crust releases large amount of energy.

The cycle of oceanic lithosphere begins at mid-ocean ridge (MOR), continues with the thickening, cooling, and moving along the earth's surface, then subduction and sinking into the mantle, and ends with the decomposition either at the 670km boundary layer between the upper and lower mantle or at the core-mantle boundary. Apart from the formation of oceanic crust at MOR (see below), this cycle will only affect mantle material that is frozen at the asthenosphere/lithosphere interface, then moves along the earth's surface in the direction of a subduction zone, and there dives down into the mantle, where in deep mantle it is finally melted again. A simple volumetric estimate of the mass of mantle material that is involved in this cycle of oceanic lithosphere (see Annex 2) shows that since the formation of the earth (ca. 4500 million years ago) nearly the entire volume of the mantle has been involved in this freezing-melting process. This documents how important the cycle of oceanic lithosphere is for the cooling of the mantle and thus the planet as a whole.

We will now describe the processes involved in the formation of the oceanic lithosphere at the MOR not in a normal geographic coordinate system (length, width, depth) but with Cartesian coordinates in 2 dimensions (depth and age instead of distance to the MOR, see Fig. 3.3).

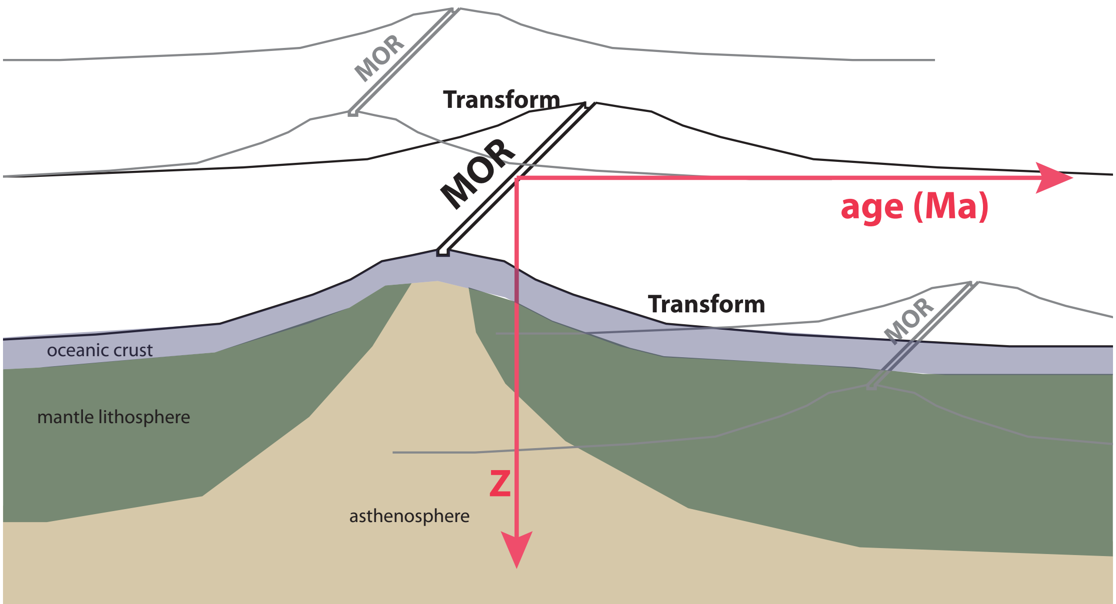


Figure 3.3: For the description of the processes involved in the formation of the oceanic lithosphere, we often use a 2D coordinate system perpendicular to the strike of the mid- ocean ridge (MOR). Whereas this view – continually displaced along the transform faults– is suitable to describe the oceanic lithosphere of an entire plate, it does not apply to the flow pattern in the mantle, where symmetrically opposing flow along the transform faults is not possible.

In all earth sciences, there only are few such clear series of observations that can be explained with a simple and unambiguous model as the data for oceanic lithosphere in the neighborhood of the MOR (see also Fig. 3.4):

$$q_{\text{heatflow}}[\text{mw/m}^2] = \frac{350}{\sqrt{\text{age [Ma]}}} \tag{3.1a}$$

$$H_{\text{water depth}}[\text{km}] = 2.5 + 3.5 \times \sqrt{\text{age[Ma]}} \tag{3.1b}$$

$$H_{\text{Lith. thickness}}[\text{km}] = 10 \times \sqrt{\text{age [Ma]}} \tag{3.1c}$$

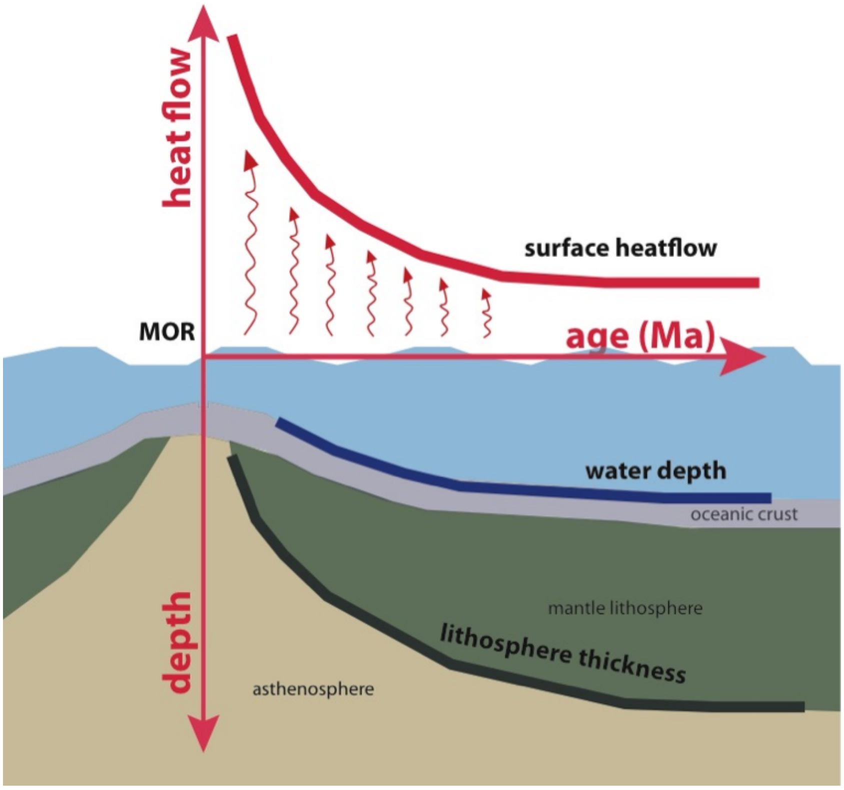


Figure 3.4: Schematic summary of the three sets of observations: surface heat flow, water depth, and thickness of the lithosphere as a function of distance from the MOR in age of the oceanic lithosphere.

The viscous asthenosphere is made of a mixture of material, which is unstable at the relatively low pressures and temperatures prevailing near the water/sediment boundary at the ocean floor. For this reason, a basaltic melt separates from the peridotite-rich asthenosphere in a few dozen km depth beneath the MOR (decompression melting). Due to its lower density and much lower viscosity the basalt rises and builds oceanic crust at the MOR (5 km to 8 km thick). Basaltic lavas contain magnetic minerals that are magnetized by the Earth's magnetic field and the direction of the prevailing field is frozen in as these minerals cool below their Curie temperatures. This thermoremanent magnetization will remain unchanged as long as the rocks are not heated above their Curie temperatures again.

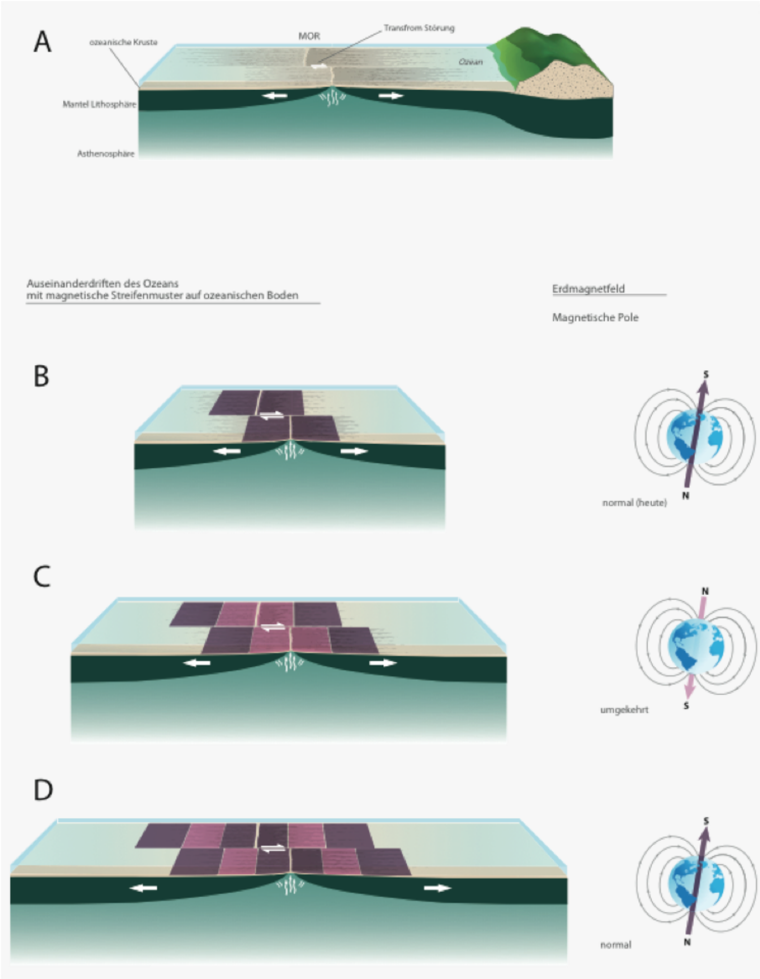


Figure 3.5: Generation of the magnetic stripe pattern of oceanic crust on both sides of the MOR in an spreading ocean (A). When cooled below the Curie temperature a thermoremanent magnetization of the basaltic oceanic crust on both sides of the MOR is frozen in, creating stripes of similarly oriented magnetization depending on the pre- vailing magnetic field. Since the magnetic field reverses polarity at irregular intervals (situation B-C-D), a pattern of stripes of alternating positively and negatively polarized magnetization are generated. These are displaced along transform faults between the MOR segments.

Because the Earth's magnetic field reverses the polarity at irregular intervals, characteristic pattern of magnetic stripes emerge on either side of the MOR (Fig. 3.5). This pattern provides information about the age and order of genesis of oceanic lithosphere. Depending on the velocity of opening at the MOR – corresponding to the rate of production of oceanic lithosphere – wider or narrower strips can arise over equal time periods. For general study of the processes involved in the formation of the oceanic lithosphere at the MOR and in its surroundings the age is therefore more suitable than the distance to the MOR, because the former is independent of the opening speed (Fig: 3.4 and equations 3.1a to 3.1c).

At the MOR, the heat flow is particularly high (Fig. 3.1 and 3.4; Equation 3.1a) because there the ca. 1250°C hot viscous asthenosphere is only covered by a 5-8 km thick lithosphere (consisting in this case only of oceanic crust) and considerable energy is released (Fig. 3.6) through the freezing of basalt melts into oceanic crust and of asthenosphere into mantle lithosphere on both sides of the MOR. Since toward the inside of the earth it gets even hotter, the energy can be released only through the lithosphere plate toward the cooler earth's surface.

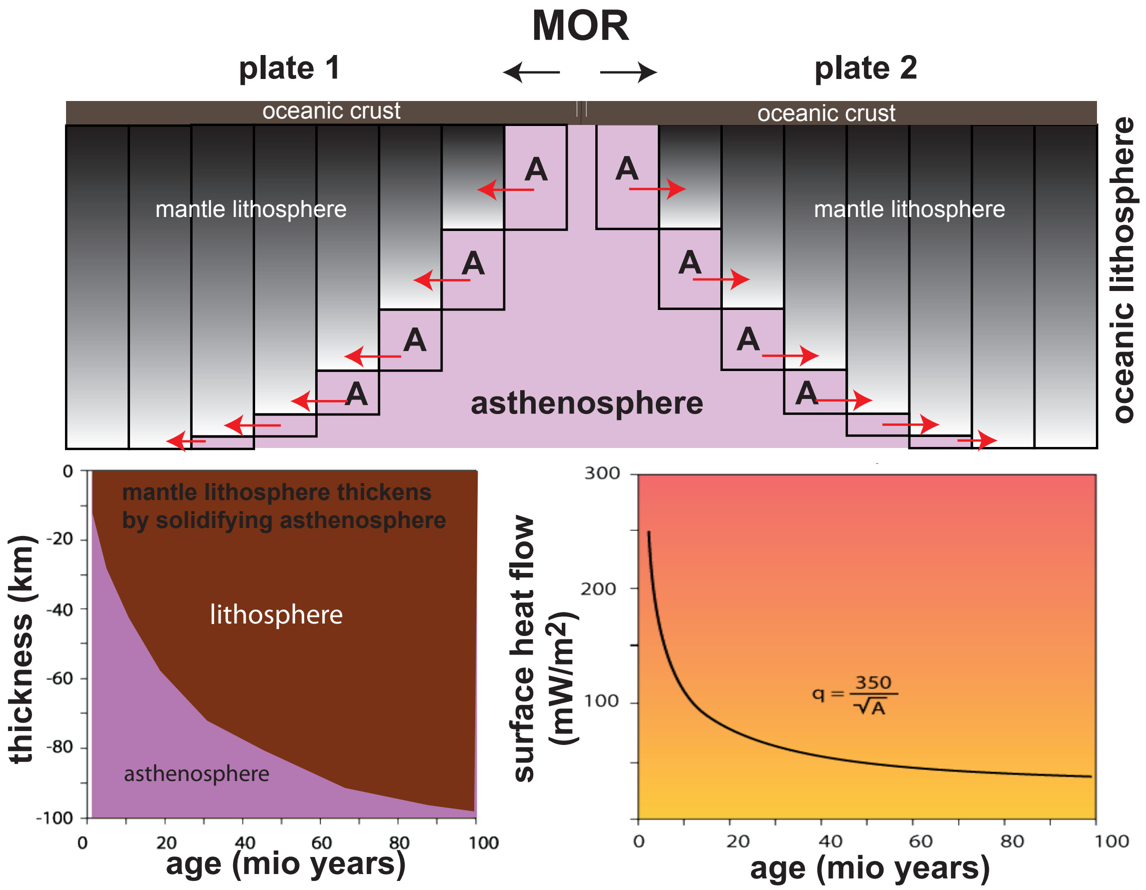


Figure 3.6: Generation of mantle lithosphere at the MOR by freezing asthenosphere material to mantle lithosphere beneath the insulating layer of oceanic crust. By this and the energy released by the subsequent cooling of the lithosphere leads to the characteristic distribution of heat flow density distribution (eq. 3.1a) and thickness of the lithosphere (eq. 3.1c) as a function of age, respectively distance to the MOR.

The thickness of the lithosphere beneath the oceans can be determined by surface wave tomography and follows the function given in equation 3.1c for the mantle lithosphere caused by freezing (at 1250-1300°C) of the asthenosphere, first beneath the oceanic crust and then laterally from the MOR on the underside of the lithosphere (Fig. 3.6). With increasing distance from the MOR and increasing age, the gain in thickness of the lithosphere per unit time decreases rapidly. With thickening of the purely conductive lithosphere, less and less asthenosphere freezes and adheres until the lithosphere has reached a thickness (today ca. 110 km) in which only just enough heat can flow through lithosphere by means of heat conduction as introduced by the mantle to the bottom of the lithosphere. From this time extra energy released by freezing of asthenospheric material can no longer be removed and the thickness of the lithosphere remains stable. Eq. 3.1c therefore applies only until the age of ca. 100 million years.

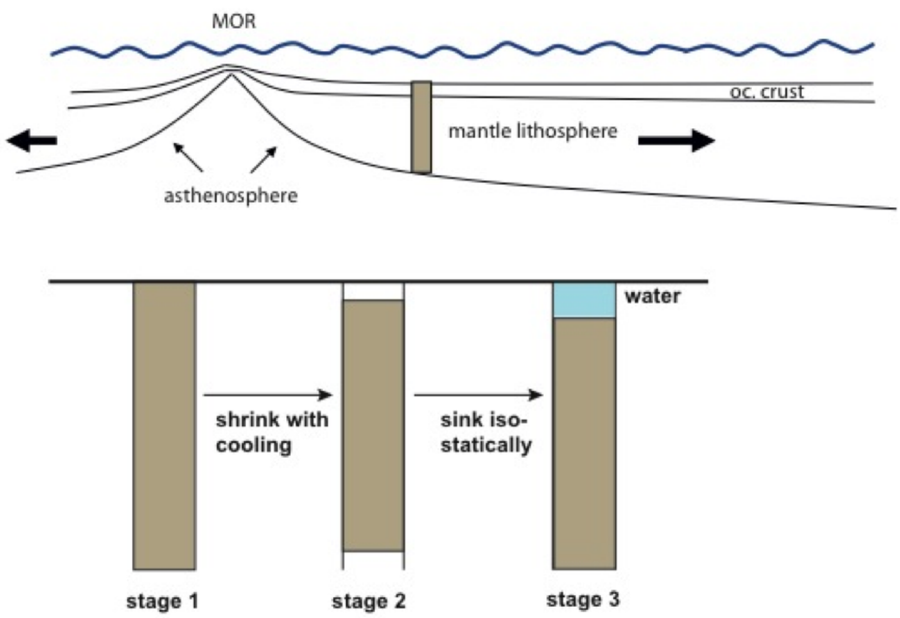


Figure 3.7: The lithosphere contracts as it cools. As a result, its density increases and thickness of a lithospheric column shrinks slightly. Since the lithosphere floats on the asthenosphere, it sinks until isostatic equilibrium is reached again. The water depth in the ocean basins is therefore approximately 5-6 km, more than twice as much with respect to above the MOR.

Easiest and most precise to measure is the depth of the oceans (see eq. 3.1b and Fig. 3.4). With the exception of Iceland, where the MOR juts above sea level due to the bulge caused by a mantle plume, we find an average water depth of 2.5 km over the MORs and an average water depth of 5-6 km in the wide ocean basins far from the MORs. The characteristic function of the water depth and age of the underlying oceanic lithosphere results from cooling processes in the newly formed lithosphere (Fig. 3.7). With increasing age, the lithosphere increases in thickness and cools, whereby the material contracts and thus increases in density. Since the lithosphere floats on the asthenosphere, with age the former subsides and the overlying ocean basin is filled with water.

The processes of the oceanic crust formation and freezing of asthenosphere material that increase the thickness of oceanic lithosphere at and in the vicinity of the MOR are only possible because the plates drift apart from each other at the ridge. This slow and in the lang-term view continuous movement of the plates and plate boundaries can be studied best using the example of the African plate. Africa is almost completely surrounded by a MOR system, in the West by the central and southern Atlantic Ocean, to the south by the South Sea, southeast by the Indian Ocean, and in the northeast by the Red Sea. The map of the age of the oceanic lithosphere (Fig. 3.8) allows the reconstruction of the position of the MOR system since its formation during the breakup of Gondwana.

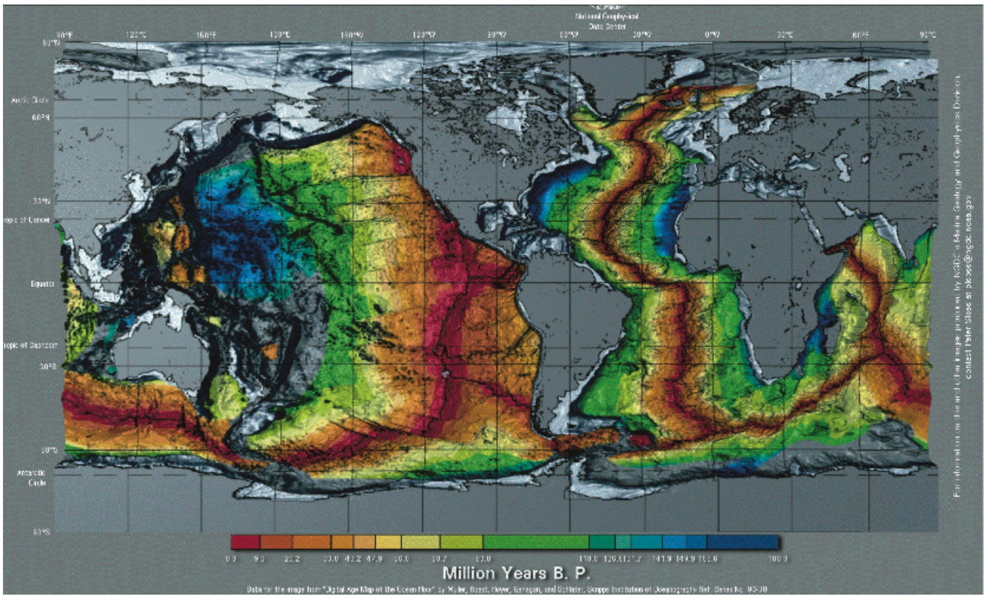


Figure 3.8: Age of oceanic lithosphere based on the pattern of magnetic stripes (Mueller et al., Scripps Institution of Oceanography Ref. Series No. 93-30).

Figure 3.9 shows the situation of the MOR system around Africa 70 million years ago in relation to the hypothesis – widespread in many older textbooks – of a divergent flow in the mantle beneath a MOR. In the case of Africa, such model implies an eastward flow on the western side of the continent (in the Atlantic) and a westward flow on the eastern side (in the Indian Ocean). This system of two opposing flows would result compressive east-west stress field in continental Africa. We observe, however, a divergence along the East African rift basin system. In addition, it would be difficult to justify what system of forces would drive the two mantle flows beneath the South-Atlantic MOR and the South-Indian MOR to drift apart with time allowing Africa to remain in the center. Finally, tomographic images indicate a diffuse, weak upward flow beneath continental Africa and no such flows beneath the MORs in the Atlantic and Indian Oceans. The observed flow may be a remnant of the superplume that was responsible for the breakup of Gondwana (Fig. 3.9B and C).

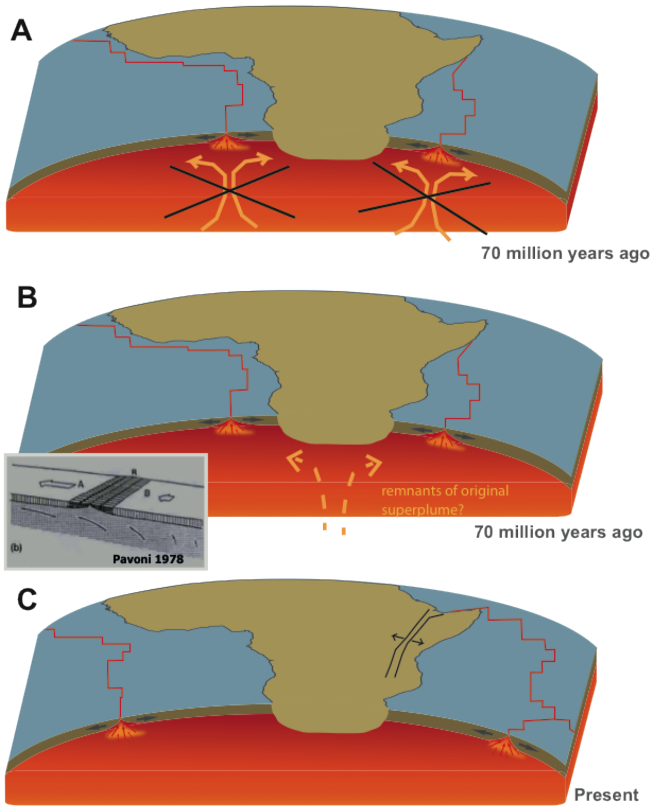


Figure 3.9: The MOR system astride Africa at 70 million years (A and B) and today (C). Deep divergent mantle flow under both MORs in the Atlantic and in the Indian Ocean (A) would produce compression within the African continent. This contradicts observations including tomography (Model B) and the presently active passive east-west separation causing the East African Rift Basin (Model C).

A divergent deep mantle flow under the MOR as a driving force for divergent plate boundaries can also be excluded because of the segmentation of the MOR and the horizontal offsets along transform faults (Fig. 3.10). Tomographic images show clearly that the increased temperature in the asthenosphere under the MOR results from local ascent of melt in the uppermost 150 km only.

The opening mechanism at the MOR is not controlled from the outside by mantle flows, but – after initiation of the drifting apart of two plates – due to local structure (Fig. 3.11) and is referred to as "ridge push." However, this force acts only in the vicinity of the MOR as long as the mantle lithosphere has not yet reached its final thickness of 100-110 km.

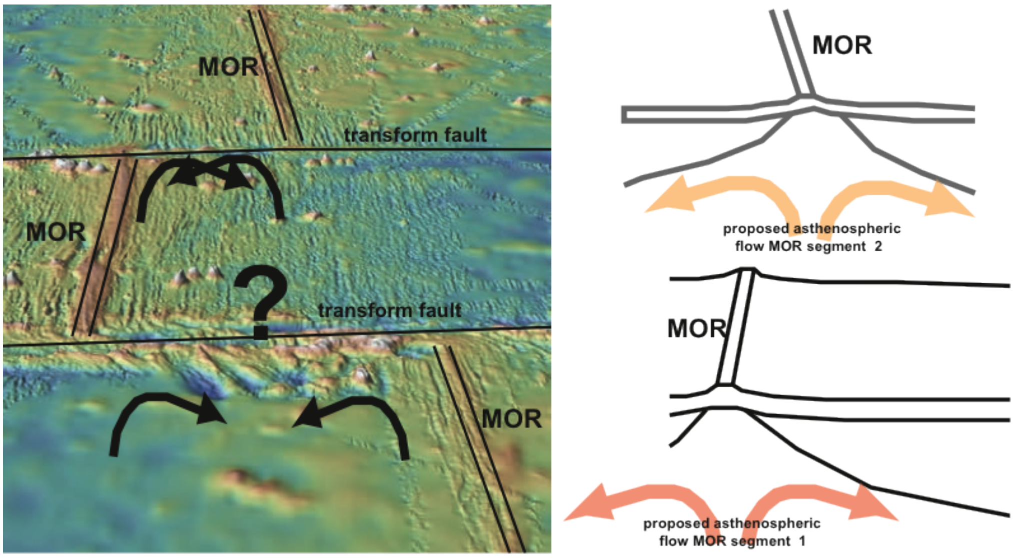


Figure 3.10: A model of deep divergent mantle flow beneath the MOR (here an example from the East Pacific Rise) that would actively push the plates apart contradicts flow laws in a viscous mantle as they imply that each transform segment would have individual flow pattern that ran counter to the flow of the neighboring transform segments.

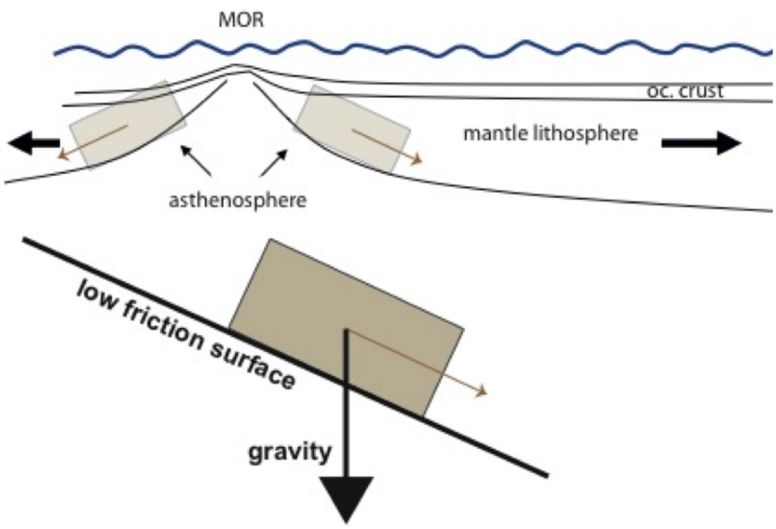


Figure 3.11: The model of gravitational sliding of the lithosphere on an inclined plane dipping away from both sides of an MOR is called "ridge push."

After less than 20 million years, the oceanic lithosphere reaches a higher average density than the asthenosphere and is therefore floating in an unstable equilibrium. The oceanic lithosphere does not sink because the plates cover the entire earth's surface and the underlying liquid asthenosphere has little possibility to reach the surface (see also Fig. 2.1). Subduction, that is, the lithosphere diving and descending into the asthenosphere occurs only under another plate and only oceanic lithosphere with density (3.28 t/m^3 for a 100km thick oceanic plate) greater than the asthenosphere ($\rho \approx$ 3.25 t/m^3) (Fig. 2.3) may do so. Continental lithosphere is less dense than the asthenosphere and cannot be subducted. Oceanic lithosphere subducts beneath another oceanic plate or continental lithosphere. When a plate consists of oceanic and continental lithosphere (see Fig. 2.5), the passive continental margin is drawn into the subduction zone at the end of the subducting oceanic lithosphere. However, the buoyancy of the continental lithosphere prevents complete subduction, so that the oceanic lithosphere slab finally tears off and sinks while the continental part of the plate remains floating at the surface.

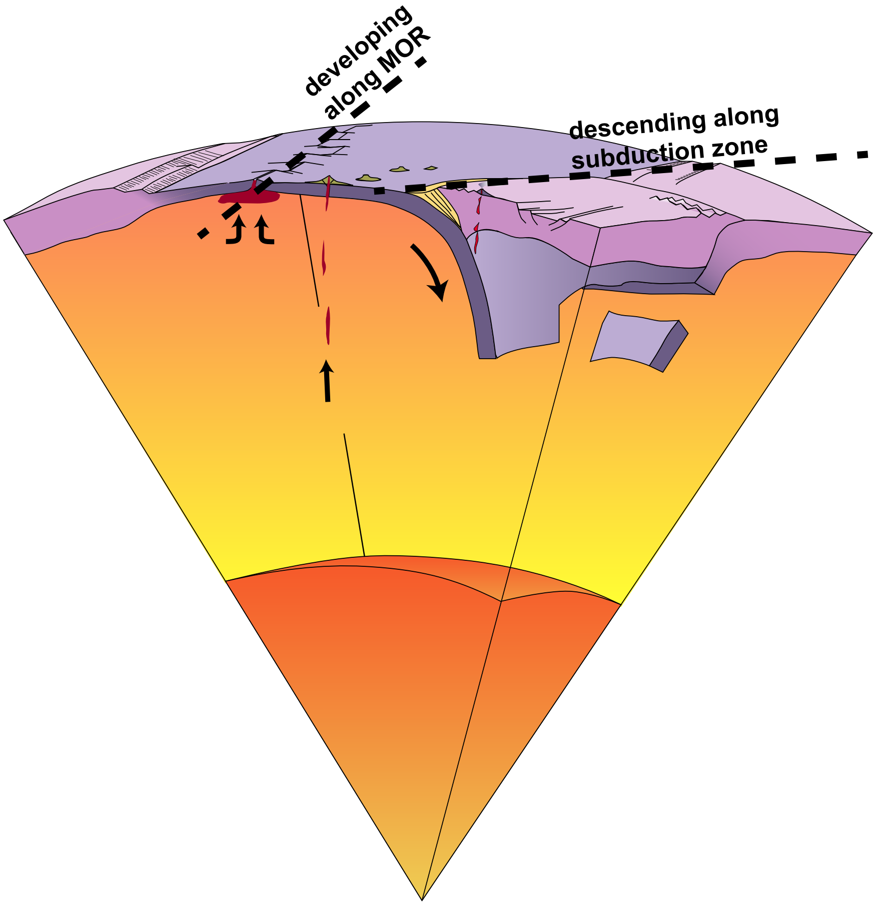


Figure 3.12: 3D perspective model of plate tectonic formation and subduction of an oceanic plate. The oceanic lithosphere subducts into the mantle like a curtain. A 2D profile section from an MOR to the subduction zone cannot adequately represent the actual forces and processes.

New oceans, that is, an MOR with a newly forming oceanic lithosphere are created when a plate breaks apart, this occurs most often with large continents. This is what happened a few million years ago when Arabia separated form eastern Africa and as a consequence created the Red Sea. Note that the new plate boundary runs along the MOR and that newly formed oceanic lithosphere on both sides of the MOR belongs to two different plates. The cycle of emergence of MOR, opening of an ocean, subduction of oceanic lithosphere, and finally closing the former ocean after all the oceanic lithosphere was subducted, is called the Wilson cycle after Canadian scientist Tuzo Wilson. Note also that this cycle occurs on a sphere and that the opening only rarely follows parallel, opposite movements (opening and closing of an ocean) and therefore the actual directions cannot be illustrated with a 2D profile section. Rather, small circles approximate the development of a MOR system, which is usually at an angle to the subduction zone of its oceanic lithosphere (see Fig. 3.12).

4 Continental lithosphere - modern and cratonic lithosphere

The continental lithosphere is our reference archive for the geological history of the earth prior to the last 200 million years ago (oldest oceanic lithosphere). At the same time the study of continents yields important clues for understanding the growth of the continents and global plate tectonics. The history of the development of continents is documented in its tectonic structure, its geological and seismic structures, in today's geothermal field, in the chemistry of the lithosphere, etc.
In essence, the continental lithosphere forms passively moved insulating plates whose outermost layer has, relative to oceanic lithosphere, a greater internal heat production due to decay of radioactive elements. There are two types of continental lithosphere, modern and cratonic (Fig. 2.3). The latter was created more than 1000 million years ago and is marked by a cratonic crust, a greater thickness, and slightly more mafic chemical composition. The cratonic lithosphere, respectively the cratonic crust, bear witness to a bygone era of plate tectonics and will not be formed anew. Cratonic lithosphere forms the core of today's continents.

Modern continental lithosphere has the same thickness (ca. 100-110 km) as mature, cooled oceanic lithosphere. This is no coincidence since the mantle lithosphere thickens by similar processes all over the globe. If just the regular heat flow from the asthenosphere can be transported through a lithosphere of ca. 100 km thickness by conduction alone, then no additional energy can be discharged. Remember that significant energy is released when the asthenosphere freezes (at about 1300°C) and consequently no more asthenosphere can be frozen onto the underside of the lithosphere. This applies more or less independent of whether the top ca. 100 km thick layer comprises oceanic lithosphere (with 5-8 km thick oceanic crust on top) or continental lithosphere (with ca. 30 km thick continental crust on top). With 150 km to 250km thickness, cratonic lithosphere (including ca. 45 km thick cratonic crust on top) is far to thick to allow any further freezing of asthenosphere. By thermal erosion through impingement of a mantle plume or the breakup of a continent due to parts of the plate separating off (Fig. 4.1), the thickness of the mantle lithosphere can be temporarily significantly reduced. However, during the subsequent cooling down to a normal geothermal gradient, the mantle lithosphere will grow back again, today attaining a thickness of only 100 km to 110 km. This means that at certain places cratonic crust is preserved within a more recent lithosphere exhibiting a thickness of ca. 110 km.

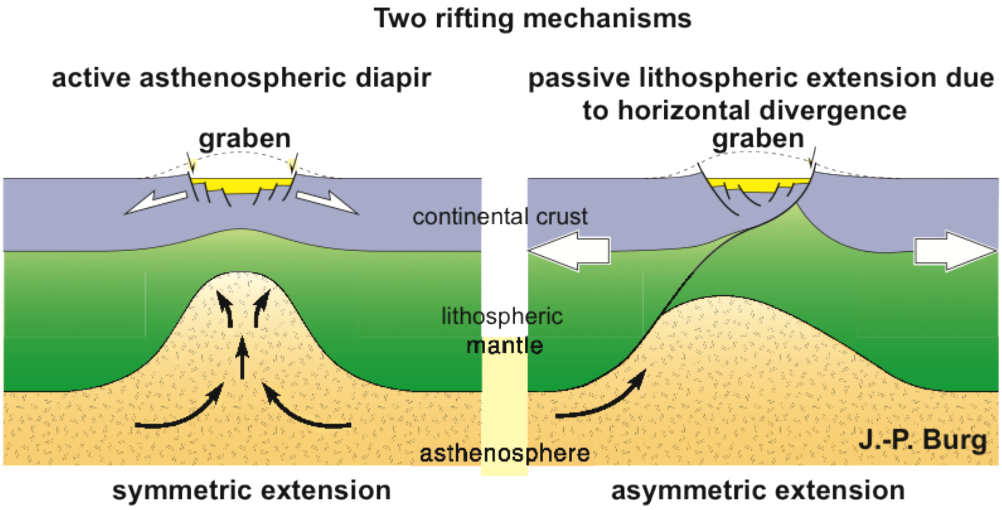


Figure 4.1: Two different mechanisms may be invoked to break a continent apart. In both cases, a rift system forms on the surface, but with differing characteristic structures. Such a rift system can evolve into complete separation of continental plates and the creation of a new ocean basin (e.g., the Red Sea). In the case of the Rhine Rift Valley the opening process came to a standstill million 40 years ago and the two bordering mountain shoulders, the Vosges and Black Forest, the Moho-bulge, and the thinned lithosphere are today observable. (Courtesy J.-P. Burg)

Young continental crust forms by continuous chemical differentiation of oceanic and continental crust and mantle material. These differentiation processes require large amounts of energy derived locally from ascending melts. Primary production sites of continental crust are the subduction zones. The subduction of water-saturated oceanic crust leads to the typical subduction volcanism, which is a surface expression of a variety of chemical differentiation processes in deeper subsurface. When oceanic lithosphere subducts beneath a continent (e.g., as is happening today under the Andes), these differentiation products can be incorporated directly at this point into the already very thick continental crust. In the case where oceanic lithosphere subducts beneath another oceanic lithosphere an island arc (e.g., the Philippines) is formed with a crust of up to 25 km thick. If such an island-arc system itself later enters a subduction zone, it will be buoyed up by its lower density, torn from the rest of the subducting plate, and stuck onto the overriding plate. In this way (almost) all more differentiated crustal material finally ends up in the continental crust, leading to the slow growth of continental lithosphere. The former separate and later accreted crustal blocks are called terranes. These are pressed onto the former edge of the overriding plate and leaving the typical image of a convergent zone, which may consist of many very different terranes (Fig. 4.3).

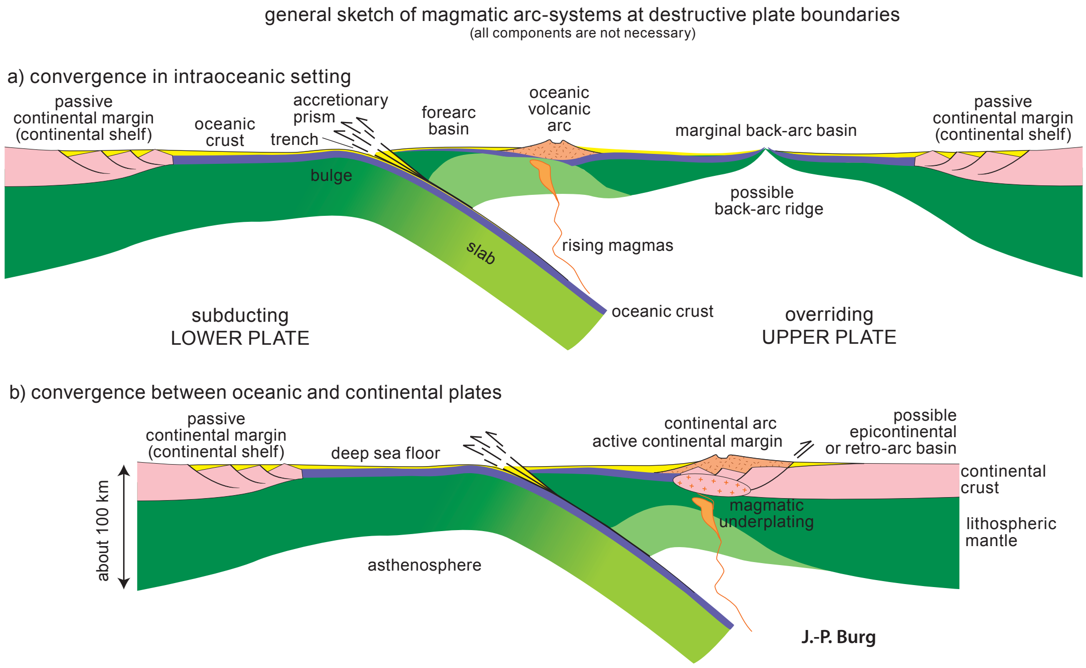


Figure 4.2: Island arcs and volcanic chains on continents are accompanying products of subduction zones. These processes form new (continental) crustal material. (Courtesy J.-P. Burg)

If an ocean is completely closed between two continents there arises a continent-continent collision (e.g., the Alps) where in the end only a few remnants of the former ocean floor are wedged between the blocks of continental crust. Such remains of former oceans are called sutures and yield – as integrated parts of a continent – information about the long-vanished ocean. In this sense, knowledge of the evolution of the Earth earlier than 200 million years ago is stored only in the continental crust and cratonic mantle lithosphere. Closure of an ocean almost always takes place in a direction different than the previous opening and often different continental plates are involved than those in the original opening of the ocean. As a result, the continents show a complex system of tectonic provinces with different ages (Fig. 4.4).

Figure 4.3: Western North America shows display a convergent zone, the typical mosaic picture of terranes that have docked here over an extend phase of subduction. (Original image by J. Cook, Woods Hole Oceanographic Institution, http://pubs.usgs.gov/gip/dynamics/Pangaea.html)

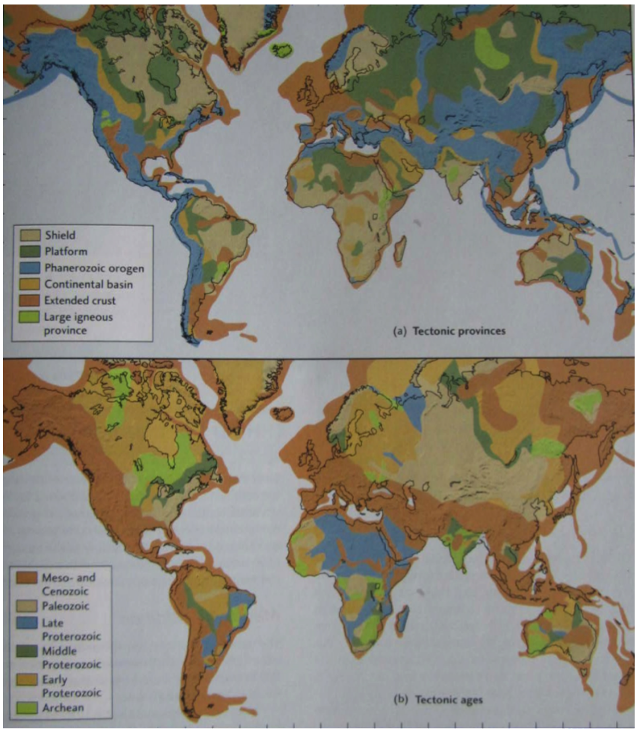


Figure 4.4: Tectonic provinces and tectonic age of the continents. (Original image by W. Mooney, USGS).

5 Plate boundaries: MOR, transform, and subduction

Due to the three basic ways in which two blocks can slide against each other, we define three tectonic systems (Fig. 5.1) and these apply also to the three types of plate boundaries.

At a MOR (Fig. 5.2), two lithospheric plates move apart and this opening leads to the formation of oceanic crust and the freezing of additional mantle lithosphere on both sides. A MOR is always divided into segments which are offset against each other by transform faults. Because of the sphericity of the earth, the length of these segments is limited to a few 100 km whereby the offsets along the transform faults may reach up to several 100 km. Due to the high temperatures and, therefore, generally smaller thicknesses and lower strengths of the plates, earthquakes in the vicinity of MORs are generally shallow with an average magnitude (max. M7). The earthquake sources are limited to the graben (rift) systems at the MOR segments and the tectonically active sections of transform faults (Fig. 5.2).

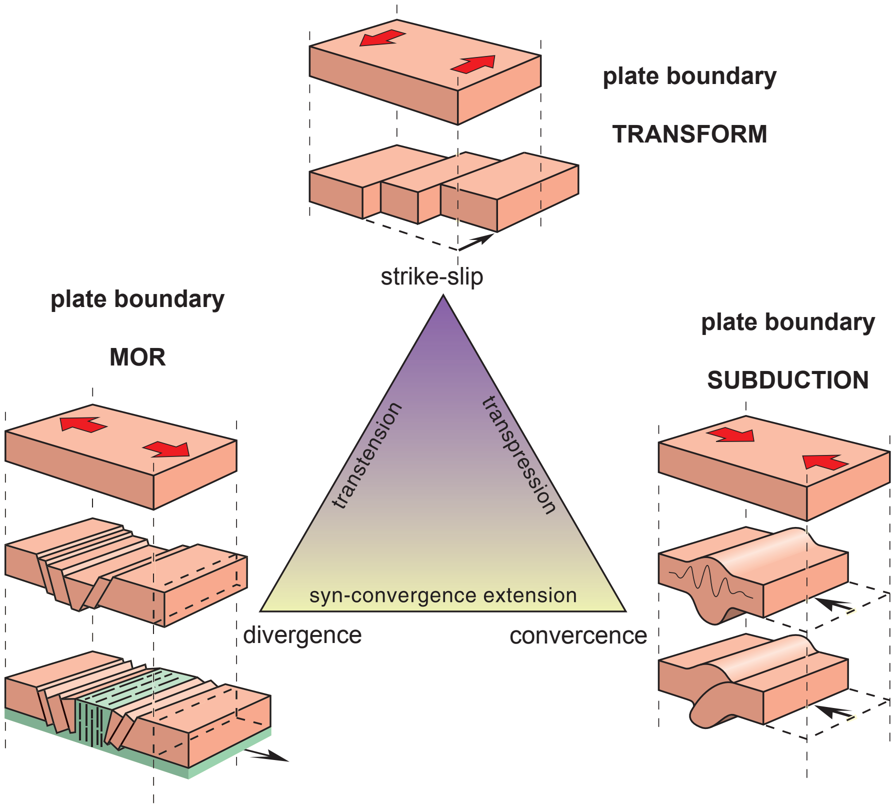


Figure 5.1: The three types of displacement of two blocks with respect to one another define three tectonic systems that are valid for both local tectonic systems as well as for major plate boundaries. Examples: strike-slip fault – North Anatolian fault zone; convergent – Alps; extensive – Rhine Graben-fault system. (Courtesy J.-P. Burg)

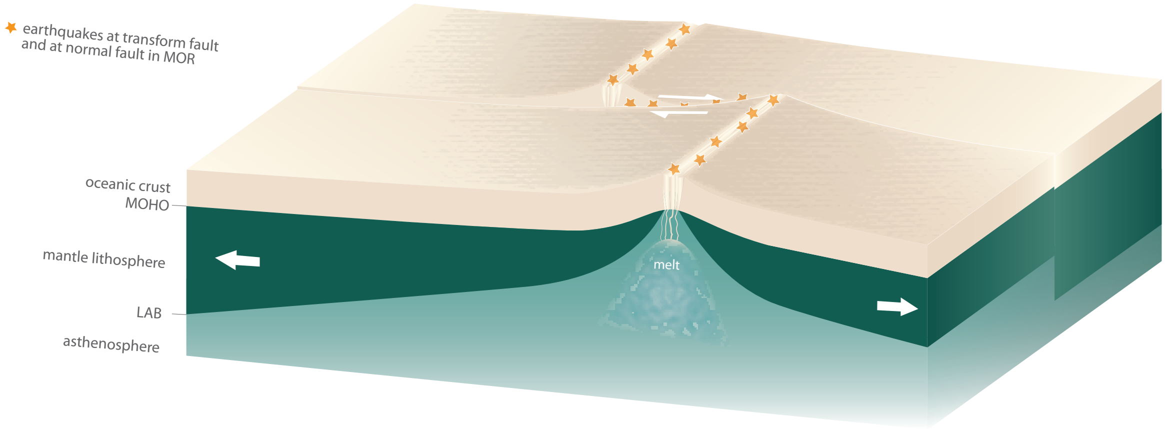


Figure 5.2: Two segments of a mid-ocean ridge (MOR), separated by a transform fault. Basaltic decompression melting in the hot asthenosphere below the MOR leads to the formation of oceanic crust. On both sides of the MOR asthenosphere freezes to mantle lithosphere, thus leading to the continual increase in surface area (see also Fig. 3.5) of the two plates.

A transform plate boundary connects either two MOR segments in a purely oceanic lithosphere environment (Fig. 5.2), or corresponds to a strike-slip fault between two lithospheric plates such as along the San Andreas fault zone (SAF, between the North American and Pacific plates, Fig. 4.3) or the North Anatolian Fault Zone (NAF, between the Eurasia and Arabia plates). In the case of the SAF, the two bordering plates along the fault zone comprise continental lithosphere with ca 25-30 km thick crust. The lower continental crust deforms in a more ductile fashion while the upper 15 km thick reacts in an elastic-brittle way and triggers earthquakes. Here, the seismogenic fracture surface reaches down to ca. 15 km depth (Fig. 5.3).

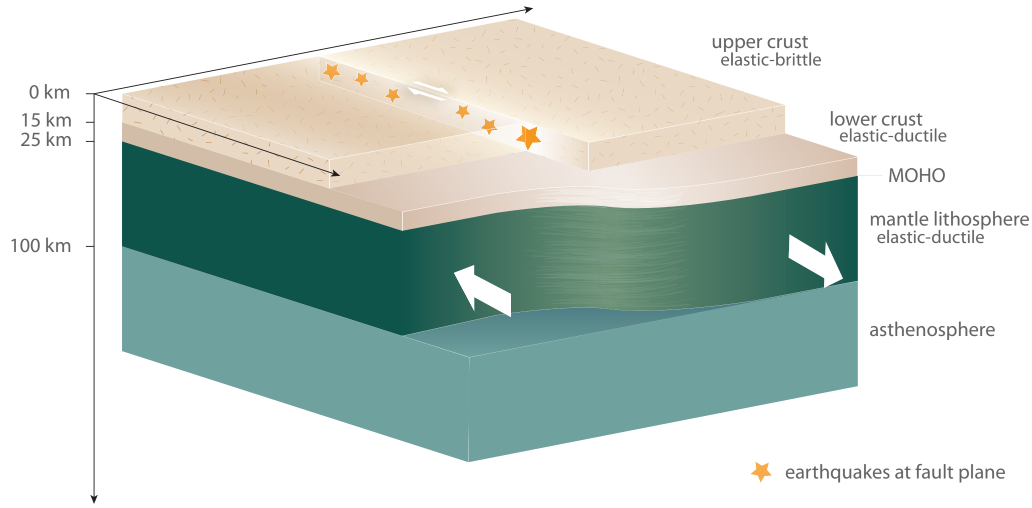


Figure 5.3: A transform plate boundary (strike-slip) where two lithospheric plates scrape past each other.

A transform plate boundary does not follow a great distance without local curves. The consequence of such changes in direction are local (pull-apart) basins and small mountainous regions (Fig. 5.4), for instance, as observed at the Dead Sea in the Jordan Rift Valley and the San Bernardino Mountains near Los Angeles.

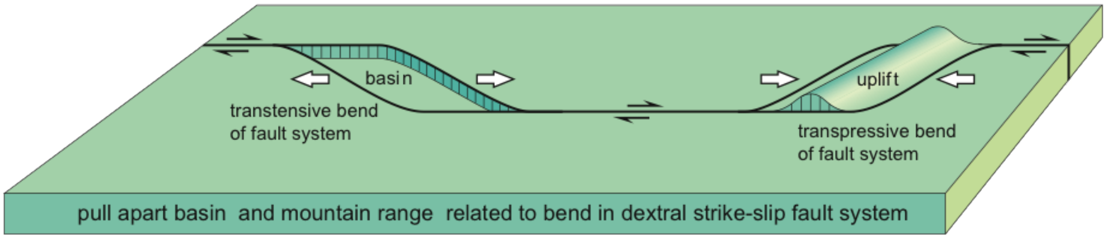


Figure 5.4: Curves in a strike-slip fault system create basins and small mountain ranges. (Courtesy J.-P. Burg)

Convergent plate boundaries are characterized by subduction of oceanic lithosphere and/or collision deformation of continental lithosphere (Fig. 5.5). Subduction of oceanic lithosphere is a run-away process meaning that once it initiates the density contrast (higher in oceanic lithosphere than the asthenosphere) drives the process onwards. Slab pull is by far the largest long-term plate tectonic force acting on lithospheric plates. It is the primary driving force responsible for the movement of the plates. This is also documented by the observation that only those plates – no matter how large or small – that have at least 30% of their total plate boundary length actively involved in subduction are moving fast.

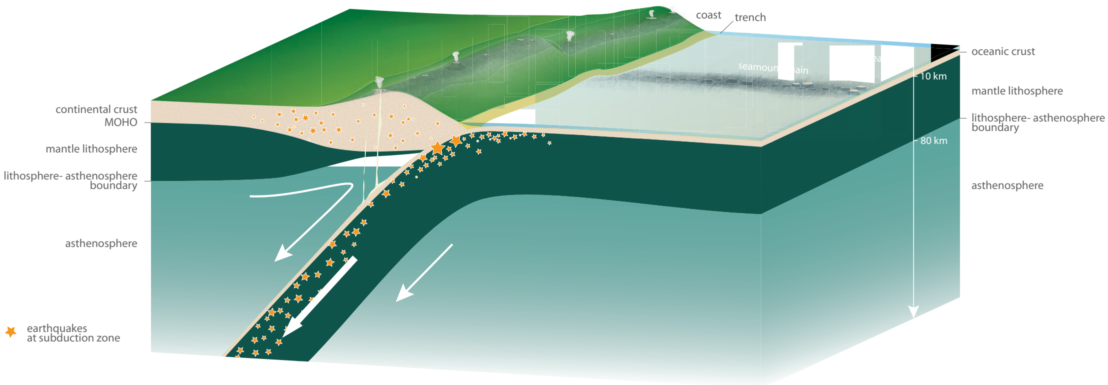
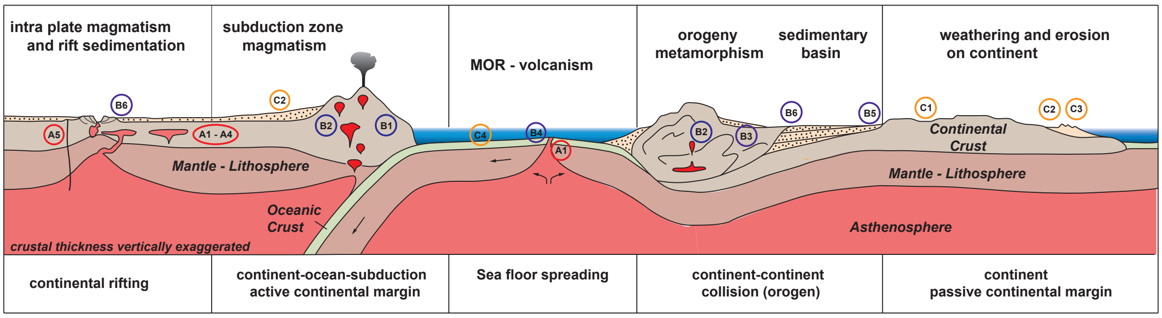


Figure 5.5: A convergent plate boundary (subduction zone). In this example, oceanic lithosphere is subducting beneath continental lithosphere along an active continental margin. The process of subduction connects the volcanism in the upper plate and the large earthquake activity in both the subducting and in the overriding plates.



courtesy C. Heinrich ETH Zürich

genetic classification of main types of ore deposits

A: magmatic ore deposits

A1. Chromitites as (ultra-)mafic cumulates in layered intrusions and ophiolites

A2. V-magnetite in mafic intrusions (V, Ti)

A3. Pegmatites (Li, Cs, Be, Nb, Ta)

A4. Ni- and PGE-sulfid deposits in mafic intrusions and flood basalts

A5. Carbonatite (REE) und kimberlite (diamond) deposits

B: Hydrothermal ore deposits

B1. Porphyr CU (Mo, Au) and epithermal Au, Ag (Hg ....) deposits

B2. Sn-W veins and greisens in granites

B3. Orogenic ('metamorphogenic') Au-quartz veins deposits

B4. Vulkanogenic massiv sulfides (Cu, Zn)

B5. Sediment-hosted (MVT, 'sedex') Pb, Zn, Cd, Cu, Co deposits

B6. Sandstone-hosted and unconformity-related U (V,F, Mo, Au, PGE) deposits

C: surface-related ore deposits

C1. Residual ore deposits: bauxit (Al), Ni-laterite deposits

C2. Alluvial placer deposits (Sn, Ta, Au, U)

C3. Beach-sand placers (Ti, Zr, REE)

C4: Manganese nodules and crusts on the ocean floor (Mn, Co, Ni, Cu)

Figure 5.6: The formation of ore deposits is closely related to plate tectonic processes. Labels A, B, C refer to major metal-transporting media and ore deposit types (Heinrich and Candela, 2014, In: Holland and Turekian (eds.), Treatise on Geochemistry, 2nd Edition, vol. 13, Oxford, Elsevier).

The largest and most dangerous earthquakes occur regularly in subduction zones along the contact surface between the plates. Through mechanical-tectonic (e.g., bending of the plate) and volcano-thermal stresses a large number of small and large earthquakes are also released in the subducting (in the so-called Benioff-Wadati zone) and in the overriding plates. During the subduction process water-saturated oceanic crust is pulled down to greater depths and exposed to higher temperatures. This lowers the melting temperatures of the material and the resulting melts mix with asthenospheric material of the above-lying mantle wedge and ascend in large quantities up into the crust of the overlying lithosphere, where they manifest themselves in various forms, one of which are the typical andesitic subduction-related volcanic chains. The Andes are an example of a mountain range formed by such a chain of active volcanoes above oceanic lithosphere subducting beneath a continent. If the subduction occurs instead beneath oceanic lithosphere, the result is an island arc (Fig. 4.2). Both forms have in common that the melting and differentiation processes lead to creation of new continental crust material (subduction factory). Subduction-related volcanism are also important in the genesis of important ore deposits (Fig. 5.6) resulting initially from water-induced partial melting of upper mantle and finally from segregation of Cu-rich fluids at upper crustal levels.

6 Mantle flow and the driving force of plate tectonics

If one lets a stone slab sink into water, it moves steeply downwards. If you imagine this occurring in a liquid of high viscosity, the sinking plate will most likely slip sideways and carry with it a large amount of adjacent sticky fluid. Hence,material flows in the viscous mantle occur either as a result of heat input from the core in combination with mantle internal heating by the decay of radioactive elements and/or as a result of sinking pieces of cold mantle lithosphere plates (Fig. 1.2). In the very young and very hot earth there were initially no plates and therefore no plate tectonics with sinking oceanic lithosphere. The earliest form of plate tectonics started with the first formation of solid plates on top of the mantle several hundred million years after the formation of the earth. The interaction of mantle flows driven by a fluid heated internally and at its bottom and by subducting slabs depends strongly on the viscosity (and thus on temperature and chemical composition) of the convecting fluid (Fig. 6.1).

At present, mantle convection (see Fig. 1.2) is dominated by two long-term and relatively continuous flow regimes:

(1) These involve moving plates and subducting oceanic lithosphere with mantle material entrained by friction. In this case, “flows” comprise both solid and liquid material. These large-scale flows are exclusively horizontal along the earth’s surface and oriented subvertically downwards in subduction. The Pacific plate serves as an example. It arises in the Southeast Pacific Ridge (MOR) from freezing local asthenospheric material. On cooling and moving ca. 10 cm/y to the northwest this oceanic plate subducts beneath various other plates along deep-sea trenches from New Zealand over Izu-Bonin to Alaska. At certain points the slab extends into the lower mantle and in other places it rests on the 670 km discontinuity.

(2) The second type of flows are mantle plumes, which often manifest themselves as hot spots at the Earth’s surface. These small-scale (local), continuous, or episodically repeating flow regimes caused by instabilities of differentiation products of subducted material as a result of heat supply either at the core-mantle boundary, or at the 670 km discontinuity (Fig. 6.2). These flows ascend primarily vertical with more-or-less (local) eddy currents and propagate radially upon reaching the bottom of the lithosphere. An example of this is the well-known hot spot plume beneath the Hawaiian-Emperor seamount chain.

Both of these long-term continuous flow regimes (Fig. 6.3) are asymmetric (i.e., there is no concurrent similar flow in the opposite direction) and non-circular (i.e., the material flow does not take place in a closed circuit, but the reflux is diffusely distributed over a large volume). Only when considered together and globally over a long time do these two mantle flow regimes yield convection analogous to that in a pot of heated water.

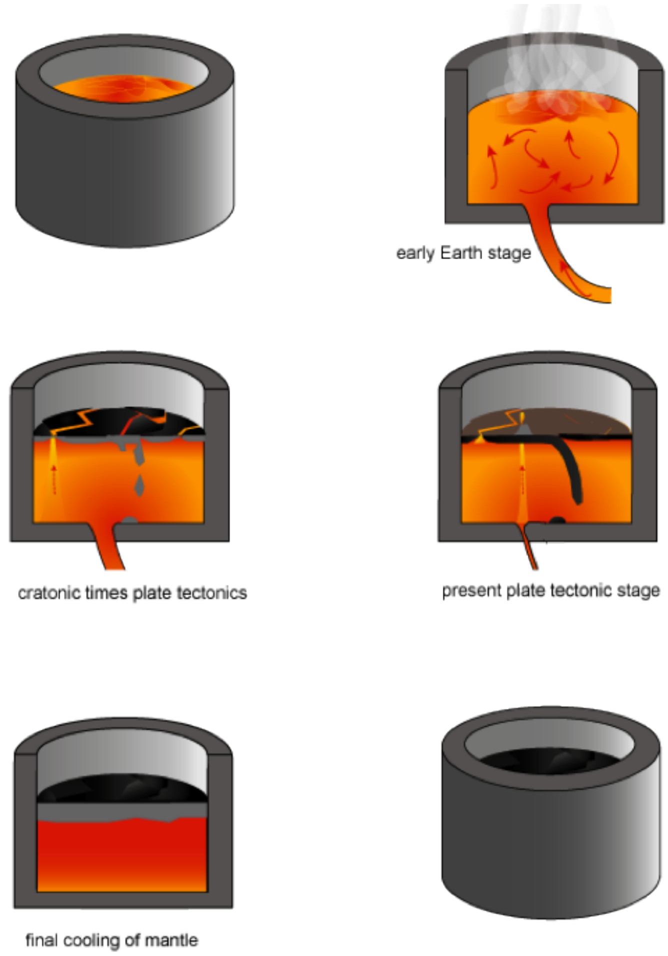


Figure 6.1: A magma lake with plates of lava in a volcano serves as a model for the system mantle-lithospheric plates of the earth. As the asthenosphere freezes at the surface to lithosphere so magma freezes to lava. Initially, the magma is too hot and active and no lava plates form. If no more heat is supplied from below, everything gradually freezes to lava. In between there is a period in which mantle plumes and plate tectonics ensure a complex dynamic equilibrium of global energy flow.

The oceanic lithosphere itself is currently by far the largest mass flow of mantle material moving with up to 10 cm/year horizontally (Pacific plate) and subvertically in subduction zones. As also seismic tomography images document, there are no symmetrically divergent asthenospheric flow cells under MORs that disperse the plates. The MORs are rather passive diverging plate boundaries (e.g., Nazca plate, which is involved in the subduction under western South America) partially supported by gravitational sliding of the newly formed lithosphere (ridge push, see Fig. 3.11).

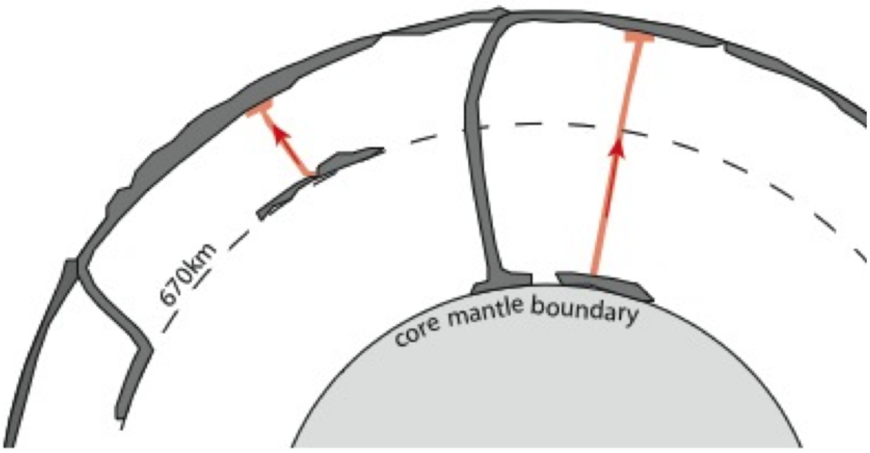


Figure 6.2: The local long-term episodic-to-continuous ascent of mantle plumes arises at the 670 km discontinuity and also at the core-mantle boundary by differentiation processes in subducted oceanic lithosphere material.

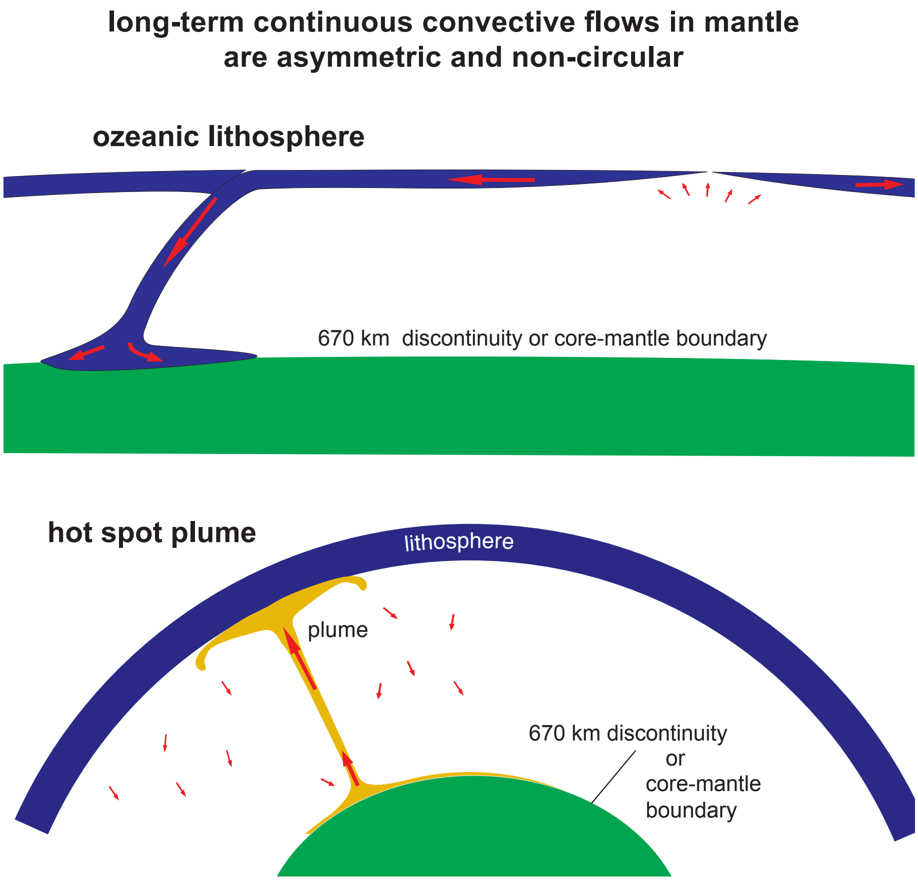


Figure 6.3: The two long-term continuous flow regimes of mantle material-oceanic lithosphere, the one originating at MORs undergoing deep subduction to the 670 km discontinuity or the core-mantle boundary, and the other driven by locally ascending mantle plumes beneath hot spots, are both asymmetric and non-circular.

The driving force for horizontal plate movements is primarily slab pull by the subducting oceanic lithosphere. Rollback subduction arises through suction forces due to the high viscosity of the asthenosphere, which can reach nearly the same magnitude as slab-pull forces. These document the emergence of “back arc basins” and the tearing off of microcontinents from the margin of an overlying lithosphere such as the case with Corsica and Sardinia.

(3) In addition to the above two continuous long-term flow regimes, there are also large-scale episodic (possibly with only large scale effect) mantel flows. These are (3a) megaplumes (episodic huge upwellings) and (3b) mantle avalanches (episodic emptying of subducted material, which have piled up at the 670 km discontinuity, into the lower mantle). Both of these occur at irregular intervals and can dominate the flow system in the mantle in the short term (Fig. 6.4).

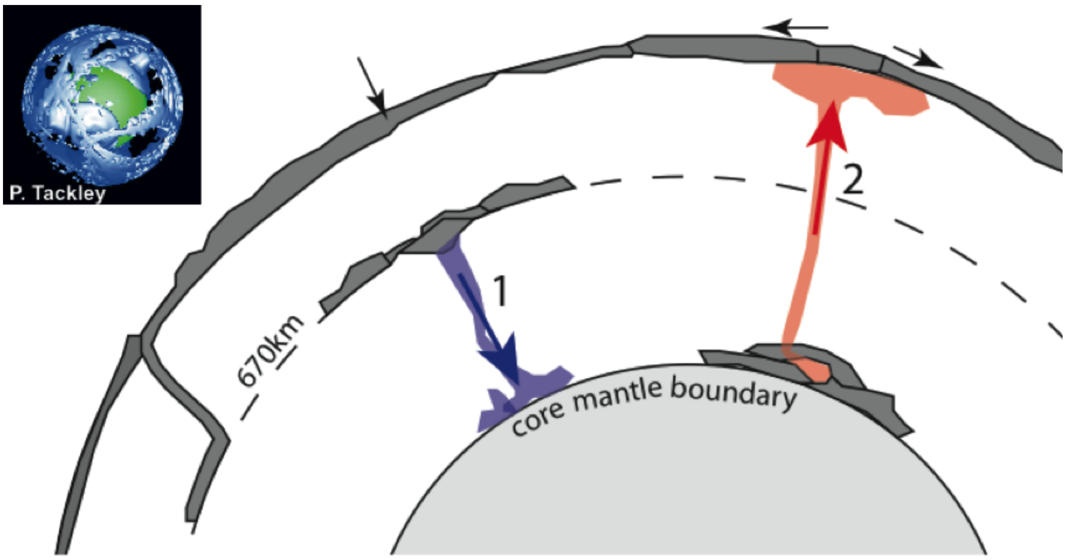


Figure 6.4: Avalanches (1) and megaplumes (2) are rare and episodic events, each with a large effect on the flow regime of the entire mantle.

Mantle flow associated with large-scale episodic mantle processes (3) have a major influence on the plate tectonic processes visible at the surface; they can even permanently alter the prevailing flow regime by breaking up continental lithosphere (including super continents), creating new oceans, and initiating new subduction zones.

After these episodic flows die down or cease altogether, the long-term regime dominated by oceanic lithosphere processes (1) and hot spot plumes (2) regain the controlling role determining the pattern of mantle flow (Fig. 6.5). These two long-term and continuous mantle flows are as a rule asymmetrical and non-circular. Strong sub-horizontal flows in the asthenosphere, which transport the overlying oceanic lithosphere by friction (the principle of MOR-symmetrical, large-scale, circular convection in the mantle), clearly make, however, the exception. This is evidenced by the radially expanding MOR system around Africa and the strong segmentation of most of MORs by transform faults. The oceanic lithosphere plates themselves are by far the most important part of the material flow along the surface of the earth and in the (cold) subducting flow regions. A rough calculation shows that the oceanic lithosphere plates, including their formation, are crucial in determining the overall heat balance of the earth's interior. About 45% of the total surface heat flow, representing approximately 56% of the heat flow from the interior of the earth is due to processes related to oceanic lithosphere.

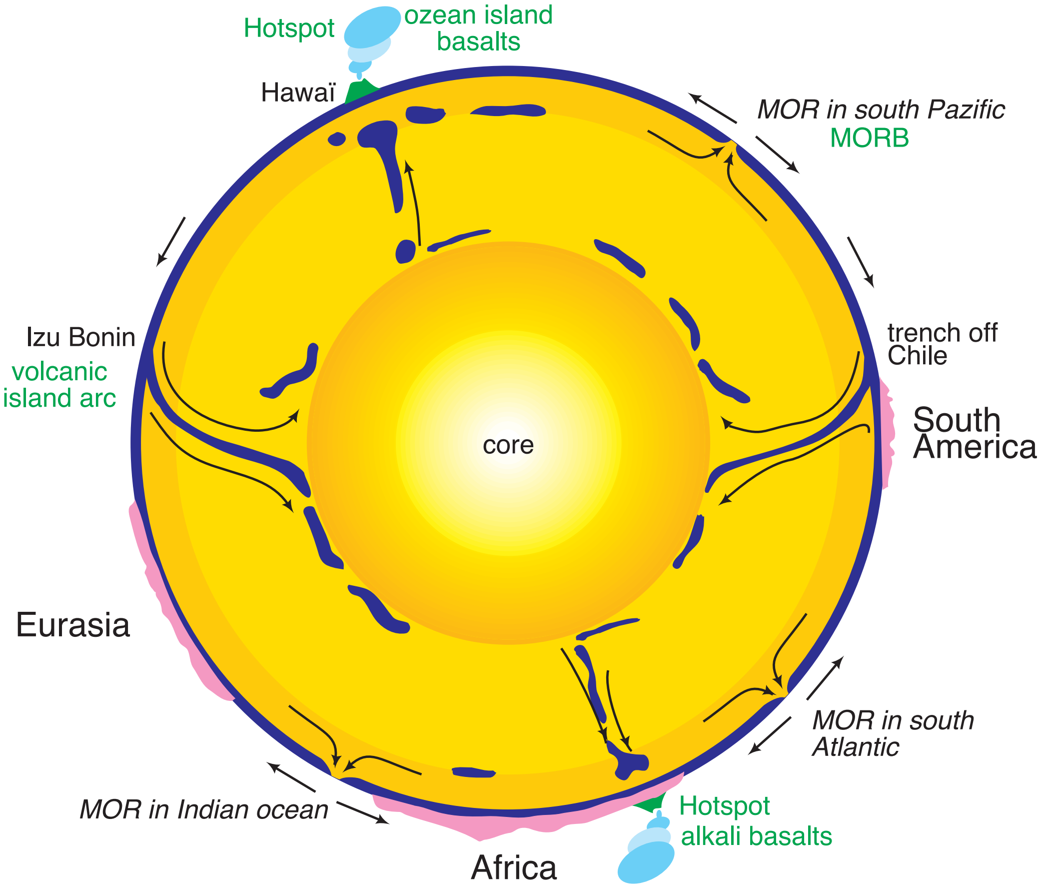


Figure 6.5: A cross section through the Earth with hypothetic flow lines in the convective mantle with lithosphere plates and their fragments along with mantle plumes.

Heat transfer from the hot interior of the earth through the mantle (including formation of the oceanic lithosphere and their recycling by subduction) to the earth's surface is ultimately responsible for the movement of the plates and thus can be called the "engine" of plate tectonics. The main force involved in mantle flows near the earth's surface result from gravity and mantle-plate interactions such as slab-pull and ridge-push. The main force in deep mantle flows (including those close to the core-mantle boundary) are the gravity and the density differences arising from temperature differences and from minor changes in the chemistry e.g., in the remnants of subducted oceanic plates. Whereas the mantle flow of oceanic lithosphere is dominated by conditions at the Earth's surface (distribution of subduction zones and the MORs), flow regimes (2) and (3) (see above) arise primarily due to local chemical and temperature differences in the deep mantle and are largely independent of near-surface structures.

Acknowledgements

For many fruitful discussions on the topic and the many years of shared teaching I thank my colleague J.-P. Burg (ETH Zürich). R. Freeman greatly improved the english text version. I thank J. Singer and S. Zukauskaité for their invaluable help with several figures.

A The 3-D heat transport equation

The 3-D temperature-dependent heat transfer equation is:

$$\frac{\partial T}{\partial t} = \underbrace{\frac{1}{\rho c_p}\nabla(k\nabla T)}_{1} - \underbrace{\mathbf{v}\nabla T}_{2} + \underbrace{\frac{1}{\rho c_p}A}_{3} + \underbrace{\mathbf{v}\frac{\alpha g T}{c_p}}_{4} + \underbrace{\ldots}_{5} \tag{A.1}$$

1. heat conduction
2. heat transport by mass transport
3. heat production
4. pressure dependence (adiabatic term)
5. any additional terms such as radiation, etc.

Whereby

T = temperature [°K or °C]

k = thermal conductivity (generally temperature and material-dependent, independent of pressure) [W/m°K]

c_p = specific heat capacity (at constant pressure) [J/kg°C] = the energy which is necessary to heat the unit mass of one degree.

ρ = density [kg/m^3]

A = radiogenic heat production (depth and material dependent) [W/m^3]

v = material velocity in [m/s]

a = coefficient of linear thermal expansion [°K^{-1}]

g = acceleration of gravity at the Earth's surface [m/s^2]

B Estimating the volume of mantle frozen as oceanic lithosphere

The radius of the core is ca. 3400 km, the radius of the earth ca. 6400 km (more precisely, 6371 km). The lithosphere is ca. 100 km thick. Using the formulas for the sphere with r = radius,

Surface of a sphere $= 4\pi r^2$

Volume of a sphere
$=(4\pi/3)r^3$

we then obtain a volume of earth of about $1.1 \times 1012\,km^3$, comprising ca. 85% mantle and ca. 15% core. A 100 km thick oceanic lithosphere over the earth's surface accounts for a volume of more than 5% of that of the mantle. The oldest known oceanic lithosphere is about 200 million years old. If we assume that the recycling of oceanic lithosphere in earlier times, due to the hotter Earth proceeded at a faster than slower rate compared with the present, then we obtain an estimate of ca. 25 cycles of generation and ca. 24 times complete subduction of oceanic lithosphere. Considering the continental crust, which does not participate in this cycle, a similar order of magnitude estimate is obtained for the total volume for frozen mantle material as for the entire mantle (25 cycles of 4% each with a mantle volume = 100%).