Earth structure and plate tectonics: Basic knowledge

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1.1 Earth Internal Structure

The Zapoliarny borehole in Russia only reaches a depth of 12.5 km; samples of deeper parts of our planet were brought up by volcanoes, but their original depth never exceeds a few hundred kilometres. Consequently, most of our knowledge of the Earth's internal composition and structure is indirect and mostly due to studies of propagation rates of seismic waves, although gravimetric data have also been used. The structure of the Earth can be schematically described as consisting of concentric shells that are from the centre to the surface (Fig. 1.1).

1.1.1 Inner Core (from 6378 to 5155 km Depth)

It is solid and mainly consists of iron with smaller amounts of nickel; its density is about 12 and its temperature at least 6000 K.

Photo 1.1. TTG sampled at 11 047.9 m depth in the Zapoliarny borehole in Kola Peninsula (Russia). This is the deepest borehole in the world until today (Photo H. Martin)

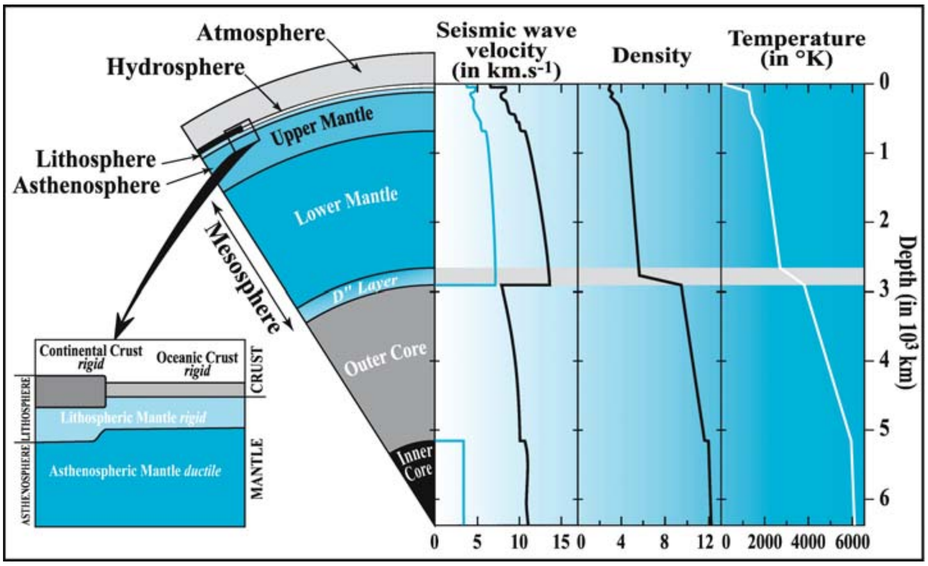

Fig. 1.1. Schematic cross section of the Earth showing its internal structure. On the right side are shown seismic wave velocity, density and temperature variations vs. depth. Shear seismic waves are in blue and compression waves in black. It must be noted that shear waves do not travel through the outer core, pointing to its liquid state. The inset on the left side illustrates the distinction that is made between mantle/crusts and asthenosphere/lithosphere pairs

1.1.2 Outer Core (from 5155 to 2891 km Depth)

It is separated from the inner core by the Lehman discontinuity at 5155 km.

Shear seismic waves (S waves) do not travel through the outer core (Fig. 1.1), which indicates that it is liquid. As for the inner core, it is composed of iron and nickel but also contains lighter element such as S, Si, O, and possibly K. From top to bottom its density ranges between 9.5 and 11.5 and its temperature between 3800 and 6000 K. As the Earth cooled it progressively crystallized at high pressure thus creating the solid inner core. During crystallization, light elements (Si, O, S, K) are rejected and remain in the liquid core.

The relative motions of the inner and outer cores as well as the convection in the outer core result in the creation of electromagnetic currents that are responsible for the Earth's magnetic field.

1.1.3 Lower Mantle (from 2891 to 670 km Depth)

It is separated from the outer core by the Gutenberg discontinuity at 2891 km.

Sometimes called the mesosphere, it consists of a silicate-rich solid, made up of perovskite $(Fe,Mg)SiO_3$ and magnesiowurstite $(Fe,Mg)O$. From top to bottom its density ranges between 4.5 and 5.5 and its temperature between 1900 and 2500 K. Velocities of shear seismic waves show that the lower mantle is solid, but on the geological timescale ($\sim 10^8$ years) it behaves as a fluid and is affected by convection.

The transition between the mantle and core consists of a 100−200 km thick heterogeneous layer called the D″ layer; through this zone temperature increases from 2500 to 3800 K. Through this thermal boundary layer, heat is only transferred by conduction and not by convection, consequently, this interface is the place where energy exchanges can take place, however, only a little matter exchange is suspected. Instabilities in this zone are considered as responsible for hot-spot magmatism. It has also been proposed that cold avalanches due to accumulation of subducted oceanic crust could create these instabilities.

1.1.4 Upper Mantle (from 670 km to 7 km Depth Under Oceans and 30 km Depth Under Continents)

It mainly consists of peridotite: a rock made up of olivine $(Fe,Mg)_2SiO_4$, orthopyroxene $(Fe,Mg)_2Si_2O_6$, clinopyroxene $Ca(Fe,Mg)Si_2O_6$ and an aluminium-rich mineral that is plagioclase $CaAl_2Si_2O_8$ (depth $< 30-40$ km), spinel $MgAl_2O_4$ ($30-40$ km $<$ depth $<$ 70 km) or garnet $Mg_3Al_2(SiO_4)_3$ (depth $>$ 70 km). Its density ranges from 3.3 to 4.5 and its temperature increases from 900−1000 K to 1900 K. It is also the place of active convection. Until now, scientific research has tried to determine if the whole mantle is affected by a single convection system or if upper and lower mantles convert independently.

The transition between lower and upper mantle is due to the transformation of olivine at low pressure into perovskite + magnesiowurstite at high pressure.

Between 100 and 200 km depth, seismic waves show a diminution of their propagation rate, this zone is called the low-velocity zone (LVZ). It is interpreted as being due to the diminution of peridotite rigidity but this zone remains solid as it allows transmission of shear seismic waves. (It must be noted that the notion of rigidity is relative and strongly depends on the timescale. For instance, the upper mantle can have an immediate rigid behaviour but is ductile when million years period is taken into account.)

1.1.5 Crusts (from 7 km Depth Under Oceans and 30 km Depth Under Continents to Surface)

It is separated from the upper mantle by the Mohorovicic (Moho) discontinuity.

There exist two different types of crusts:

Oceanic Crust

Its average composition is that of a basalt, it is thin ($\sim$ 7 km), has a density of 3.1 and its age never exceeds 180 Ma. It is generated in oceanic ridge systems by partial melting of the upper mantle and recycled into the mantle in subduction zones.

Continental Crust
It has an average thickness of $\sim$ 30 km but under mountains, it can be 70 km thick. It is highly heterogeneous with an average composition of granodiorite. When compared with mantle it has a relatively low density (2.75) such that it has high buoyancy and is almost not recycled into the mantle. This is why continental crust as old as 4.04 Ga can be found. Most of the new continental crust is generated in subduction-zone environments.

1.1.6 Hydrosphere

It consists of surface water: ocean, lakes, rivers, ice caps, clouds, etc. It strongly interacts with surface rocks.

1.1.7 Atmosphere

This is the gaseous envelope of the Earth, it can extend up to 500 km in altitude. This interface between the surface of the Earth and the extraterrestrial domain plays the role of energy filter.

Both the hydrosphere and atmosphere are the places where life developed.

1.1.8 Lithosphere and Asthenosphere

The distinction between core, mantle and crust is based on compositional differences. However, for the same chemical and mineralogical composition, depending, for instance, on temperature, rheological properties can be completely different. For instance, viscosity is strongly temperature dependent, it decreases when temperature increases. Consequently the uppermost part of the mantle that is cold has high viscosity and a rigid behaviour even at long timescale, whereas the deeper and hotter parts have a lower viscosity resulting in a ductile behaviour on the geological timescale ($\sim$ 100 Ma). Consequently, in the mantle–crusts system two rheological shells are distinguished:

Lithosphere
It consists of the crust plus the rigid part of the upper mantle, its average thickness is about 100 km under oceans and 150−200 km under continents. It has a rigid behaviour and is not affected by internal convection; it corresponds to a domain where heat is transferred by conduction.

Asthenosphere
It corresponds to the deeper and hotter parts of the mantle where peridotite has a ductile behaviour and where heat is dissipated by convection. It is generally considered that the lithosphere–asthenosphere limit corresponds to the 800°C isotherm; it roughly begins with the LVZ. As it cools the asthenosphere is transformed into lithosphere.

1.2 Plate Tectonics

Since the 1960s, the concept of sea-floor spreading has been established. As the surface of the Earth can be considered as constant, the oceanic crust, which is progressively generated in rift systems, must return and be recycled into the mantle (subduction zone). Then plate tectonics was born and this concept progressively evolved and refined.

1.2.1 Plates on the Surface of the Earth

The surface of the Earth is divided into lithospheric plates made up of crust (oceanic or continental) and with the rigid part of the upper mantle; they move over the convecting asthenospheric mantle. Today 12 major plates are recognized on the surface of the Earth but several small (micro-plates) also exist (Table 1.1; Fig. 1.2).

Table 1.1. Name and surface of the main lithospheric plates

Plate name	Area ($10^6\,km^2$)	Plate name	Area ($10^6\,km^2$)
Pacific	103.28	South-American	43.62
African	78.00	Nazca	15.63
North American	75.89	Philippine	5.45
Eurasian	67.81	Arabian	5.01
Antarctic	60.92	Caribbean	3.32
Indo-Australian	59.07	Cocos	2.86

1.2.2 Margin Definitions

Some of these plates contain both oceanic and continental crust (i.e. Eurasian or African), others are exclusively oceanic (i.e. Pacific or Nazca). Most of the seismic and volcanic activity of our planet is concentrated at plate margins. Two kinds of continent–ocean limit must be distinguished:

– ***Passive margin:*** both oceanic and continental crust belong to the same plate (limit between Europe and East North Atlantic), there is no seismic or volcanic activity;
– ***Active margin:*** oceanic and continental crusts do not belong to the same plate; generally the oceanic crust disappears under continental crust giving rise to an important seismic and volcanic activity.

There exist 3 main kinds of boundary between lithospheric plates:

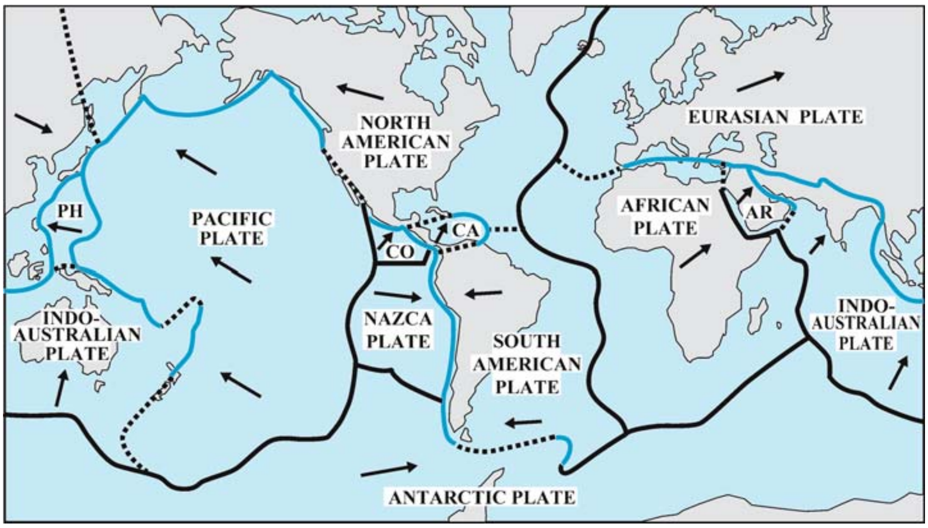

Fig. 1.2. Schematic map of the surface of the Earth showing the 12 main lithospheric plates. *Black heavy line* = divergent margin; *blue heavy line* = convergent margin; *heavy dotted line* = transform margin. Arrows show the main direction of plate motion. CO = Coco plate; CA = Caribbean plate; PH = Philippine plate; AR = Arabian plate

- ***Divergent margin:*** also called constructive margin, this is where new material resulting from upper-mantle melting is added to the oceanic lithosphere; the two plates pull away from each other. They correspond to the midoceanic ridges that form a 80 000 km long submarine relief.
- ***Convergent margin:*** also called destructive margin. This is where the oceanic crust plunges under another plate and sinks into the mantle (subduction). Subduction can occur under either continental or oceanic lithosphere. Because of its low density and of its correlated high buoyancy, continental crust is almost never subducted deep into the mantle. Consequently when both plates contain a continent, the two continents collide and form a mountain chain as, for instance, the subduction of the Indian plate under the Eurasian plate led to the collision of continents and gave rise to the Himalaya Mountains.
- ***Transform margin:*** also called conservative margin, there, crust is neither produced nor destroyed as the plates slide horizontally past each other. The San Andreas Fault perfectly illustrates this type of contact between Pacific and North-American plates.

1.2.3 Divergent Margin

Midoceanic ridge

The rate of oceanic spreading ranges between 2 and 17 cm yr^{-1}.

Divergent margins are located over the ascending branch of a convection cell (Fig. 1.3). Due to the low thermal diffusion rates, solid upper mantle peridotite ascents adiabatically. Figure 1.4 shows that during adiabatic ascent a dry ($H_2O \sim 0.1\%$) peridotite will crosscut its dry solidus temperature (temperature where dry peridotite begins to melt) at low pressure (in the domain of stability of spinel or even of plagioclase) thus giving rise to a basaltic magma whose crystallization generates new oceanic crust.

As this magmatism is generally submarine, ocean water penetrates into the fractures of the newly formed oceanic crust.

During its descent into the crust water progressively warms up, thus creating a convective water circulation into the oceanic crust. This water will not only exchange energy with basalts but it will also chemically interact with it. This is well exemplified by the black or white smokers that correspond to the escape of warm ($> 300\,°C$) hydrothermal water rich in sulfurs dissolved from the basalts. Another important effect of water is that a basalt initially composed of anhydrous minerals (olivine $(Fe,Mg)_2SiO_4$, orthopyroxene $(Fe,Mg)_2Si_2O_6$, clinopyroxene $Ca(Fe,Mg)Si_2O_6$ and plagioclase $CaAl_2Si_2O_8$) will be transformed into a hydrous mineral-bearing rock (talc $(Fe,Mg)_3Si_4O_{10}(OH)_2$;

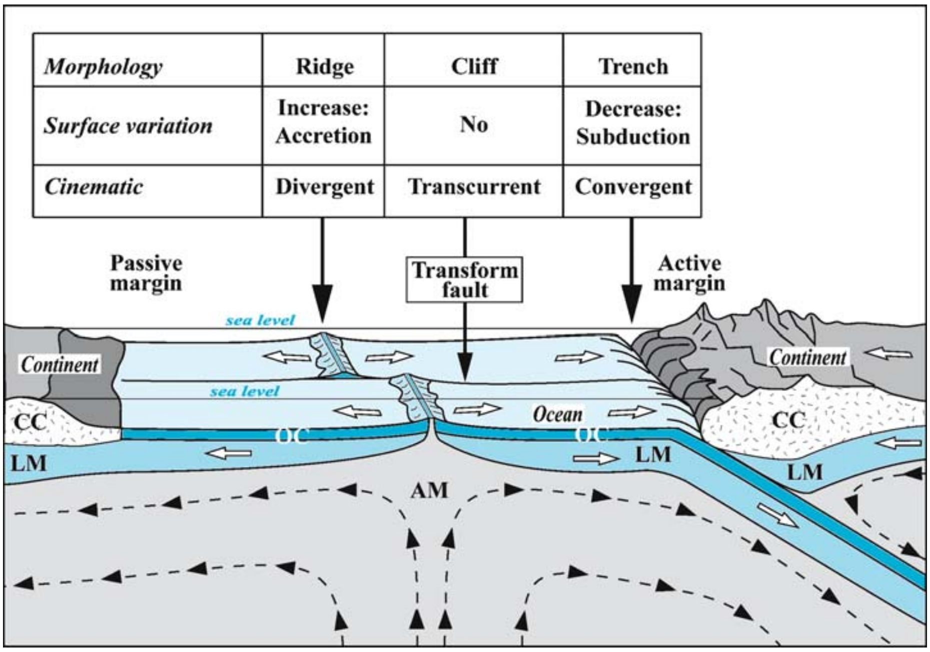

Fig. 1.3. Schematic block diagram showing different possible relationships between plates as well as the terminology generally used. CC = continental crust; OC = oceanic crust; LM = lithospheric mantle; AM = asthenospheric mantle. *Black arrows* and *dotted lines* show convective displacement into the upper mantle

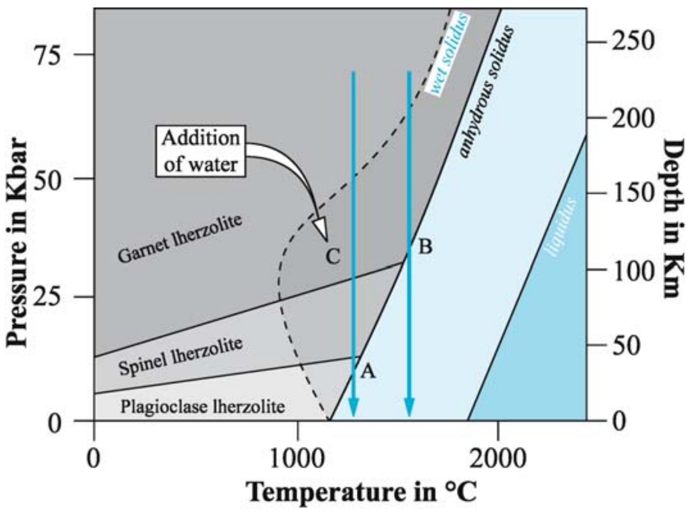

Fig. 1.4. Pressure vs. temperature diagram showing the domains of stability of garnet, spinel and plagioclase lherzolite of the lithospheric mantle. *Heavy lines* represent solidus and liquidus of dry mantle, *dotted line* is the solidus of hydrous mantle lherzolite. In ridge environments mantle melts by adiabatic decompression, the dry *solidus curve* is cross-cut at shallow depth (**A**). Melting in hot-spot environments is also due do adiabatic decompression but as mantle has a deeper origin it is hotter and melting takes place at greater depth (**B**). In subduction zones, mantle melts by addition of water due to dehydration of the subducted oceanic crust, water significantly lowers the solidus temperature (**C**)

serpentine $(Fe,Mg)_6Si_4O_{10}(OH)_8$, amphibole $Ca_2(Fe,Mg)_5Si_8O_{22}(OH)_2$, chlorite $Fe_3Mg_3AlSi_3O_{10}(OH)_8$). The oceanic crust that contained about 0.3% water when it emplaced will contain between 1.5 and 5% water after hydrothermal alteration.

In the course of time the newly formed oceanic crust moves far from the ridge and progressively cools. Therfore, near the ridge the lithosphere was warm and consequently thin whereas as it cools the lithosphere becomes progressively thicker.

Continental rift

A continental rift is a zone of continental lithosphere where extensional deformation takes place and that generally evolves towards continent break-up and ocean birth (Fig. 1.5). The process starts with the ascent of hot mantle material under continental crust. As this material is hot and buoyant, it causes the doming of the overlying plate. The warmed lithosphere transforms into asthenosphere. Doming of continental crust provokes its stretch and thinning due to its split into large blocks limited by faults, thus resulting in a collapsed valley called rift. As thinning proceeds, crust begins to be pulled apart, to extend and as in oceanic ridge, adiabatic decompression of mantle peridotite leads to its melting and to the genesis of basaltic magma. Progressive and successive injection of basalt in the central part of the rift contributes to pushing plates apart, the opened space being immediately filled with magma. This mechanism results in the genesis of oceanic crust, whose surface increases as continents separate (Fig. 1.5).

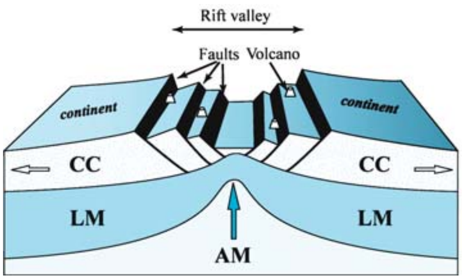

Fig. 1.5. Schematic diagrams showing how continental crust breaks (*left*) and how ocean birth occurs (*right*). CC = continental crust; OC = oceanic crust; LM = lithospheric mantle; AM = asthenospheric mantle

1.2.4 Convergent Margin

Subduction zone

To compensate for new material formed in ridges, old oceanic lithosphere has to be destroyed which is realised in subduction zones (Fig. 1.2). As it cools, the density of oceanic lithosphere increases and consequently, its buoyancy decreases. When its density becomes greater than that of the underlying asthenosphere, then oceanic lithosphere can spontaneously sink into the mantle. In this case as it is old, oceanic lithosphere is thick and subduction angle is high (Fig. 1.6).

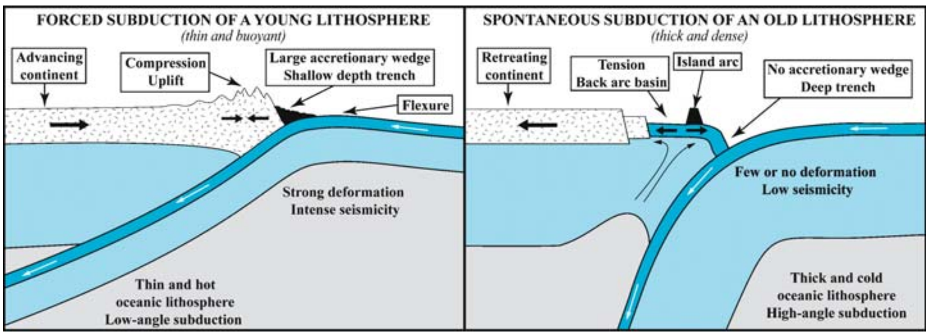

Fig. 1.6. Schematic cross sections showing the contrasted characteristics of both spontaneous and forced subduction

This geometry induces a deep trench and generally the associated seismicity is moderate (i.e. Marianna, Japan).

When strain constraints are too high young oceanic lithosphere can be forced to subduct, even if its density is lower than the asthenosphere one. In this situation (Fig. 1.5) the lithosphere is thin and angle of subduction is low (it can even be almost flat). As oceanic lithosphere resists subduction, seismic activity is generally very important. All sediments emplaced on the ocean floor do not enter into subduction and form an accretionary wedge in a relatively shallow trench. Thus, the continental crust is compressed and can form mountains (i.e. Andes).

Contrary to ridge systems, subduction corresponds to descending branches of convection cells (Fig. 1.2); consequently dry mantle cannot melt by adiabatic decompression. As it enters into subduction, the hydrated oceanic crust progressively dehydrates (metamorphism) such that it becomes unable to melt. The liberated fluids rise up into the overlying dry mantle wedge. Interaction of water with dry peridotite significantly lowers its solidus temperature (Fig. 1.6) such that it can melt. Fluids released by subducted oceanic crust not only lower mantle solidus temperature, but they also transport dissolved elements (U, K, Rb, etc.) that modify the mantle peridotite composition (metasomatism). The generated magmas are andesitic or granodioritic in composition, they form new (juvenile) continental crust. The descending residual oceanic crust accumulates at

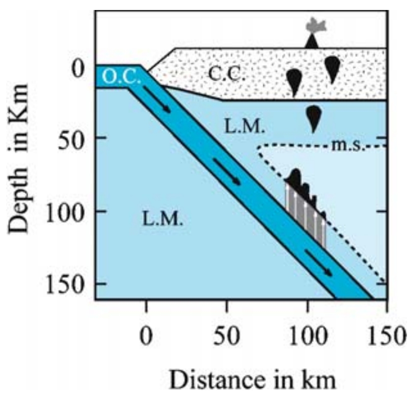

Fig. 1.7. Schematic cross section of a subduction zone showing the process of magma generation: as it enters into the mantle, the subducted oceanic crust dehydrates. Liberated fluids rise-up through the overlying mantle wedge, which becomes progressively rehydrated, this lowers its solidus temperature and induces its partial melting (Fig. 1.3). O.C. = oceanic crust; C.C. = continental crust; L.M. = lithospheric mantle; m.s. = lithospheric mantle hydrous solidus (the *pale blue part* of the mantle wedge corresponds to the part whose temperature is greater than the hydrous solidus one: it can melt if water is available); *black domains* correspond to place where magma is generated and emplaced; whereas *grey area* with *white arrows* represents place where fluids pass through

the 660-km seismic discontinuity. When stored oceanic crust exceeds a threshold it suddenly sinks into the mantle as a cold avalanche that reaches the mantle–core boundary (D″ layer). In some exceptional situations generally related to subduction of very young oceanic lithosphere, oceanic crust can melt thus generating magmas called adakites that also will contribute to form juvenile continental crust.

Collision zone
When the plate implied in subduction carries a continent, this latter, due to its high buoyancy and contrarily to oceanic lithosphere, does not significantly sink into the mantle; which results in continental collision (Fig. 1.8). The overstacking of continental crust segments significantly thickens continental lithosphere and generates mountain chains. Part of the continental crust transported to depth, can melt and generates granitic magmas. As these latter are produced by melting of the continental crust they do not constitute new addition to this crust whose volume remains unchanged: this mechanism is called recycling.

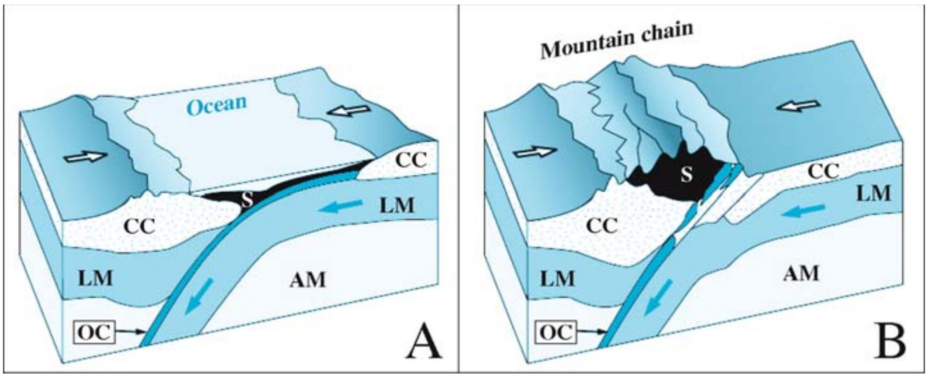

Fig. 1.8. Schematic cross sections showing how subduction and closure of an ocean (**A**) can evolve towards the collision of two continental crusts. This process, still active in the Alps and the Himalayas, generates mountain chains and thickens continental crust (**B**). O.C. = oceanic crust; S. = sediments; C.C. = continental crust; L.M. = lithospheric mantle; A.M. asthenospheric mantle

1.2.5 Hot Spots

In some places volcanism occurs in the middle of plates (Fig. 1.9) it is called hot-spot magmatism (Hawaii, La Réunion, etc.). These magmas are considered as being generated by the ascent of hot lower mantle (mesosphere) diapir, possibly coming from the core mantle boundary (D″ layer). The source of these diapirs appears as immobile on the scale of several tens of millions of years. Lithospheric

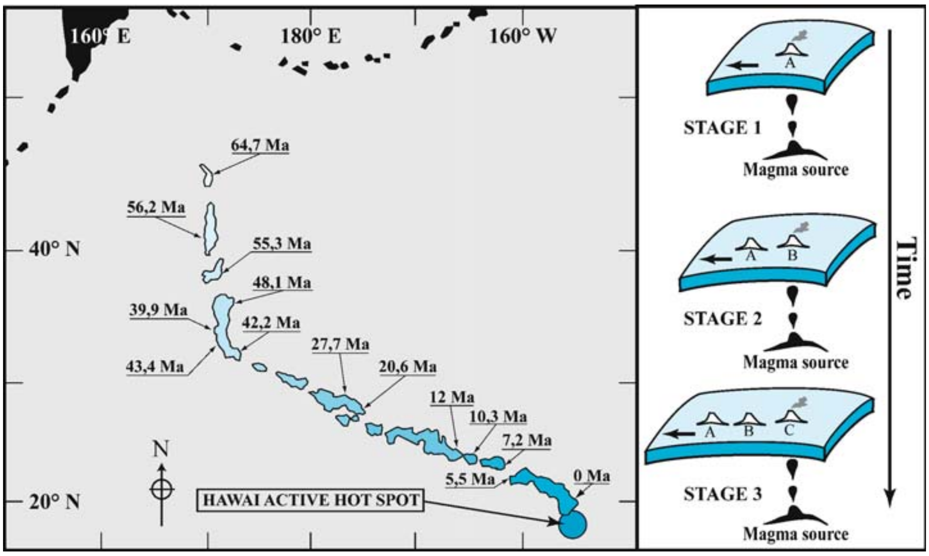

Fig. 1.9. Map of Hawaii-Emperor chain showing that the age of volcanoes regularly decreases towards Hawaii that remains the unique active edifice. *Inset*, on the *right*, schematically illustrates hot-spot island genesis. Magma source is immobile and deep seated, the lithospheric plate is perforated by ascending magma and its motion generates the volcanic chain

plates move over these fixed spots and are perforated by ascending magma, thus this motion generates chains of volcanic island (i.e. Emperor Chain, Hawaii; Fig. 1.9). Melting of the mantle is due to its adiabatic ascent, but as diapir is hotter than in the case of the midocean ridge, melting occurs at greater depth in asthenosphere between 100 and 200 km (Fig. 1.3).

Hot-spot activity can also generate large igneous provinces (LIP), which corresponds to the emplacement of huge amounts of basaltic magma either in continental (i.e. Deccan trapps) or oceanic environments (i.e. Ontong-Java, Kerguelen oceanic plateaus). Their formation requires that enormous volumes of hot mantle rise up until the base of the lithosphere. They are generally attributed to melting in the head of large starting mantle plumes (diapirs).

Recently, three great mesosphere domains (Hawaiian, Tristan and Icelandic) were recognized based on their hot-spot characteristics. They are called "mesoplates", their upper surface corresponds to the asthenosphere-mesosphere frontier and they are limited by the deep subduction zones that extend below 200 km.

1.2.6 Wilson Cycle

In 1968 Wilson developed the concept of cyclicity of oceanic spreading: Ocean begins its life by breaking a continent, it extends through a ridge system and disappears in subduction or collision; this evolution is known as the Wilson cycle. On the other hand, collision between two independent continents assembles them and leads to the formation of a single and greater new continent. Succession of subductions and continental collisions gives rise to supercontinents. The supercontinent cycle involves formation of a supercontinent from smaller continental blocks followed by fragmentation, new ocean opening and again assembly of a new supercontinent. Such a cycle lasts about 500 Ma. For instance, about 245 Ma ago, almost all present-day continents were joined and formed a supercontinent called Pangea. At 245 Ma the Atlantic Ocean began to open and India started to migrate from Madagascar towards Asia: a new supercycle started. Continent breaking continues today, for example in the Red Sea and in the East African rift.

The breakup of an existing supercontinent is due to hot-spot activity and also to the thermal shield effect due to continental crust that acts as a screen that makes internal heat release more difficult. In both cases, heat accumulates under continental crust and provokes its thickening that leads to its breakup.

1.2.7 Energy for Plate Tectonics

All the energy necessary to allow plate tectonics to work is the internal energy of the Earth. Today, this energy has been evaluated at 42 TW (42×10^{12} W). Its main source is the decay of long-lived radiogenic elements (^{235}U, ^{238}U, ^{232}Th and ^{40}K). They are mostly concentrated in the continental crust that produces about 20% of the heat of the Earth. They are less concentrated in the mantle, but due to its greater volume, it contributes about 55% of the heat budget. The core does not contain significant amounts of radiogenic elements. However, crystallization of the outer core produces latent heat ($\sim$ 10%). The last 15% of released heat corresponds to fossil accretion energy. When it formed, due to accretion energy, the Earth was hot, since 4.55 Ga it slowly cooled (200 to 250 °C since 4.0 Ga).

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