

CONTINENTS ON THE MOVE

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Summary

Earth's outer form changes continuously over the geological time associated with dramatic alterations of the atmosphere, hydrosphere, and biosphere. Our planet Earth is a living organism. It breathes, develops, and gets slowly old. Mantle convection stirs the entire mantle, generating lithospheric plate movements, and transporting great quantities of heat from Earth's interior to the surface, where the heat radiates into space. Plate movements are the most important motors of the geological events taking place in the lithosphere, and the changing of the global distribution of continents and oceans. As plates move and continents travel over them, ocean basins open and close associated with the recycling of the ocean floor. At the spreading centers, a new oceanic crust is created, while at the subduction zones, the cool, old oceanic crust sinks into Earth's interior where it is destroyed molten. The continents grow progressively, but with different rates in their evolutionary history, getting heavier and consequently more slow-moving. The continents periodically join in a single supercontinent, colliding with each other. Each supercontinent lasts for a few million years, and then breaks again into smaller fragments which will form the new continents, each riding away from the others on its own tectonic plate.

1. Introduction

Geological time is divided into smaller portions for convenience. Based on relative age determinations, large geologic time units are split into smaller intervals. From the largest time portions to the smaller, geologic time is subdivided into eons, eras, periods,

and epochs. This chronological arrangement is called the geologic time scale, and is used by the geologist to date the studied rocks and tectonic events (Figure 1).

Our planet is approximately 4.6 billion years old (see *Solar System*). The earlier eons—the Proterozoic, Archean, and Hadean, which comprise 4 billion years of geologic time—are not subdivided at all, even though together they constitute a time interval almost eight times as long as the Phanerozoic. Furthermore, we know from astronomical observations that the planet Earth will possibly stop to exist after five billion years from today. This means that overall, Earth’s life could be estimated as approximately ten billion years.

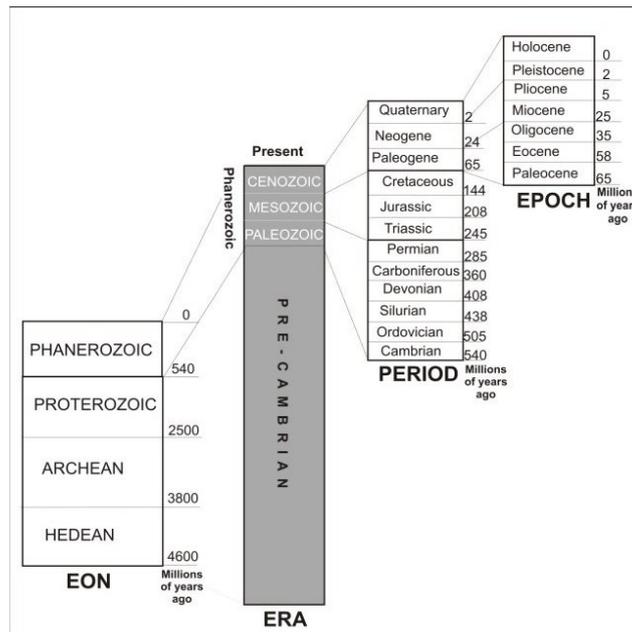


Figure 1. The geologic time scale

The appearance and evolution of human life on Earth has taken place during about the past three million years: this is a very short time in Earth’s whole life.

In nearly 4.6 billion years of Earth’s life, its outer form has changed many times, and will progressively change again in the future as a result of the effect of successive geological processes. Magmatism, metamorphism, sedimentation, tectonics, and erosion merge and will continue to merge the structure of our planet Earth.

The Greek philosopher Hiraklitos (535–475 BC) first felt this continuous transformation, and said, “*Τα πάντα ρει*” (all is flux). Many years later, the German researcher Alfred Wegener formulated his theory about continental drift, and the break up of the Pangea supercontinent.

The aim of this article is to describe the global tectonic movements that have taken place during the geological evolution of Earth, causing continuous changes to the surface fabric and climate.

2. Earth's Structure Today

2.1 Earth's Layers

From the top to the bottom, our Earth is composed mainly of three different layers: the crust, the mantle, and the core. Figure 2 illustrates the structure of Earth's layers, and the relationships between the boundary conditions of the lithospheric plates.

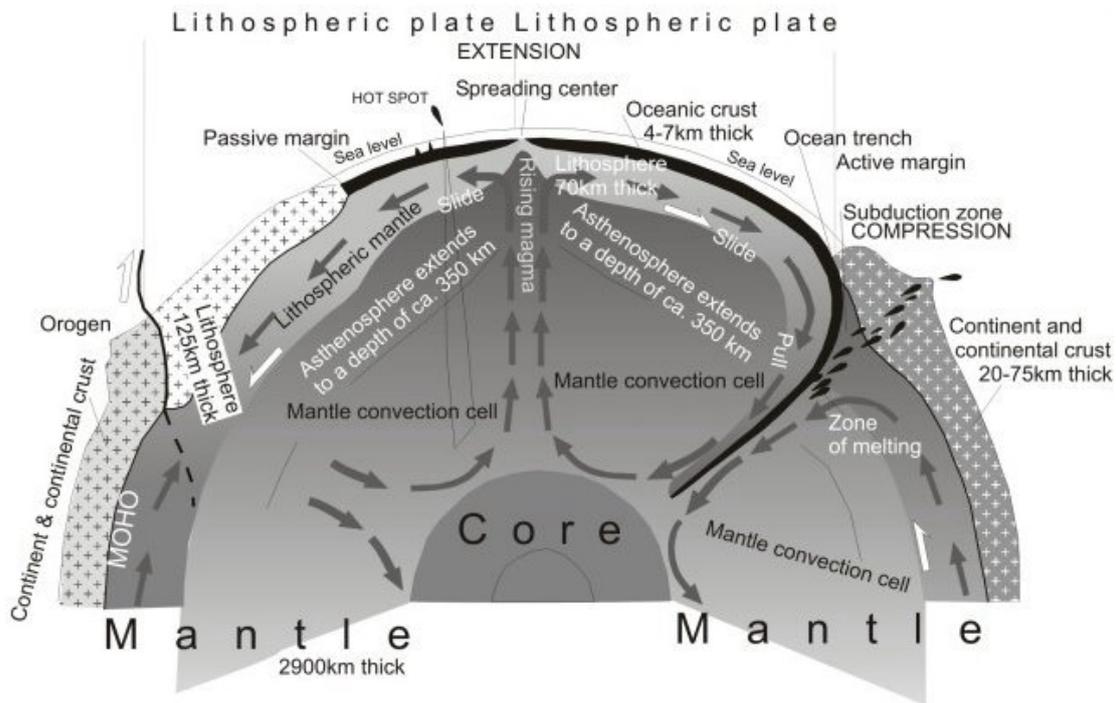


Figure 2. The structure of Earth's layers (modified, after Thompson and Turk, 1999)

2.1.1 The Crust

The uppermost and thinnest layer is called the crust. It is relatively cool, and consists of hard, solid rocks. We distinguish between oceanic crust and continental crust. The oceanic crust is composed mostly of dark, dense basalt, and extends beneath the oceans, occupying mainly the floor of the oceans (see *Oceanic Crust*). The oceanic crust is between 4 km and 7 km thick. The continents, with their marginal parts (active or passive) that extend beneath the sea level, comprise the continental crust (see *Continental Crust*). The continental crust is much thicker than the oceanic crust. Its thickness ranges (approximately) from 20 km to 70 km. The continental crust is composed primarily of light colored, less dense granite.

2.1.2 The Mantle

The mantle lies directly below the crust. It is almost 2900 km thick, and is composed of ultramafic rocks (Peridotite, Harzburgite, etc.) of which the mineralogy varies with depth (see *Mantle and Core*). Mantle temperature and pressure increase with depth.

These P-T changes cause the strength of mantle rock to vary with depth, and thus create sublayering within the mantle. We distinguish between the upper and the lower mantle.

The upper mantle is further divided from top to the bottom into (a) the lithospheric mantle, (b) the asthenosphere, and (c) the remainder upper mantle.

- a) The uppermost mantle is relatively cool, and consequently consists of hard, strong rock with the same mechanical behavior as the crust. The lithospheric mantle and the crust constitute together the lithosphere, which forms the outer hard part of the Earth. The lithospheric thickness varies from approximately 75 km beneath the ocean basin to approximately 125 km under the continents. The boundary between the base of the crust and the top of the more rigid mantle is called the Moho-discontinuity.
- b) The asthenosphere extends beneath the lithosphere to a depth of approximately 350 km. In contrast to the lithosphere, the asthenosphere is hot, weak, and plastic, so the strong, hard lithosphere floats on the soft, plastic rock of the asthenosphere.
- c) At the base of the asthenosphere, increasing pressure overwhelms the effect of rising temperature, and the strength of the mantle increases again. Therefore, the remaining lower part of the upper mantle, of about 300 km width, and the upper part of the lower mantle are mechanically strong. Despite the strength of these mantle parts, high temperature makes the entire mantle plastic and able to flow slowly, over geologic time.

Recent evidence suggests that the lowermost mantle at the core boundary is partly molten.

2.1.3 The Core

The core is the innermost of the Earth's layers (see *Mantle and Core*). It extends from a depth of 2900 km to the center of the Earth, forming a sphere with a radius of about 3470 km. The core is composed largely of iron and nickel.

The outer core is molten because of the high temperature in that region: near its center the core's temperature is about 6000°C. Despite this high temperature, the extreme pressure that dominates in the inner core compresses it to a solid state.

2.2 Tectonic Plates and Their Movements

(See *Tectonic Processes*.) The lithosphere itself is broken today into seven large and several smaller segments called tectonic plates or lithospheric plates (Figure 3). They are not plain structures, but portions of a global surface.

The lithospheric plates glide slowly and horizontally over the asthenosphere at rates ranging from less than 1cm/a to about 16 cm/a (cm/yr). A single plate can carry both oceanic and continental crust. Only the Pacific plate is composed entirely of oceanic crust. Continents travel on the lithospheric plates as they move over the asthenosphere, like passengers.

A plate boundary is a discontinuity in the lithosphere that separates two lithospheric plates. A plate margin is tectonically active, in contrast to the interior of the plate, which is normally stable. All significant processes of Earth's dynamics are placed at the plate boundary: this is where mountain ranges or ocean basins build up. Magmatic eruptions, metamorphism, and earthquakes also take place here.

Neighboring plates can move relative to each other in three main ways. We distinguish between a *divergent boundary*, *convergent boundary*, and *transform boundary* (Figures 2 and 3). Each plate boundary type is characterized by its typical functions and tectonic activity.

Combinations can take place between divergent or convergent boundaries and a transform boundary. In these cases of oblique plate divergence or convergence, we distinguish the difference between *transtensional* and *transpressional tectonics*, respectively.

- (a) At a divergent plate boundary, also called “spreading center” or “rift zone,” two plates spread apart from one another. The underlying asthenosphere then oozes upward to fill the gap between the separating plates. As the asthenosphere rises between the separating plates, some of it melts to form magma, due to decreasing pressure conditions rather than temperature change. Most of the magma rises to the Earth's surface or the sea floor, where it cools to form new oceanic crust. So, a continuous and intense volcanism takes place along the divergent plate boundaries. As the asthenosphere rises between two separating plates, it gains mechanical strength by cooling, and therefore transforms into new strong lithosphere. In this way, new lithosphere is continuously formed at a divergent boundary, and its thickness increases as it moves away from the spreading center. The lithosphere can be as little as 10 km to 15 km thick at a spreading center (see *Processes that Form Magma*).
- (i) Horizontal extension dominates at a divergent boundary.
- (ii) The two separated plates at their divergent boundary may be composed of oceanic crust or continental crust. Respectively, we speak of ocean rifting or continental rifting, although continental rifting will often be developed into an oceanic rifting over some time, for example, the continental rifting in East Africa. In this case, if the rifting continues, eastern Africa will be separated from the main body of the African continent, and a new ocean basin will open between the separating portions of Africa.
- (iii) Mainly shallow earthquakes are related to the spreading processes.
- b) *At a convergent plate boundary*, the plates move towards each other. Horizontal compression dominates at a convergent plate boundary. Where two plates converge, the denser one sinks into the mantle beneath the other, while the less dense plate remains floating on the asthenosphere. This sinking process is called subduction (Figures 2 and 3). The subducted plate can reach a depth of approximately 700 km into the mantle.

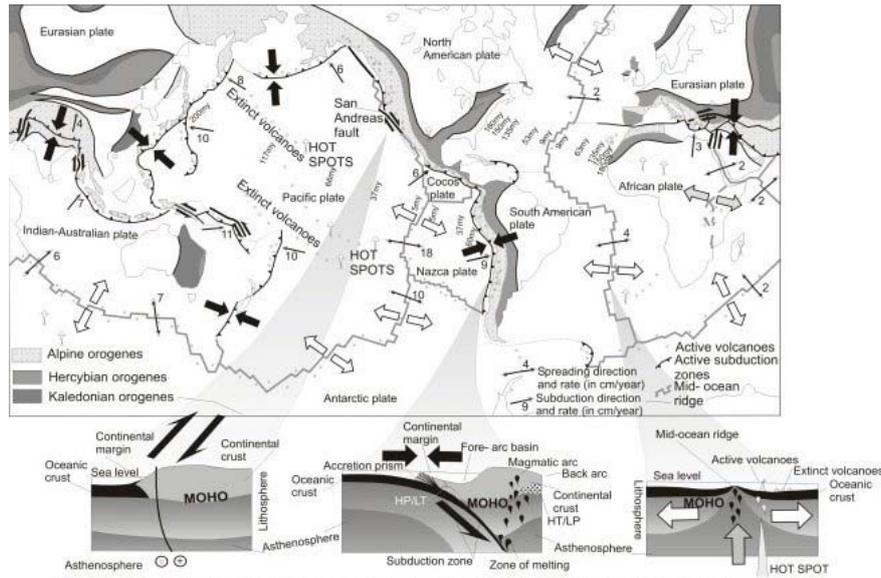


Figure 3. The distribution of the main lithospheric plates and their boundary relationships with the associated geological processes

- (i) In most cases only oceanic lithosphere can sink in a subduction zone into the mantle, since it is denser than the thick continental crust. However, in certain cases, small amounts of continental crust may be forced into the mantle in a subduction zone.
- (ii) In general the rate at which old lithosphere sinks into the mantle at subduction zones is equal to the rate at which new lithosphere forms at a spreading center. Therefore, Earth's dynamics maintain a global balance between the creation of new lithosphere and the destruction of old lithosphere. The oldest sea-floor rocks on Earth are only about 200 million years old, because oceanic crust continuously recycles into the mantle at subduction zones (Figure 3). Rocks as old as about 4 billion years are found on continents, because subduction consumes little continental crust. Plate convergence processes are in general associated with mountain chain building, intense crustal subhorizontal shortening, and vertical thickening. Deep earthquakes, rising magma, volcanism and metamorphism are immediately related to the plate convergence processes.

Transform plate boundaries develop where two plates slide horizontally past one another as they move in opposite directions (Figure 3). This type of boundary can occur in both ocean and continent. Along the transform plate boundaries, neither new lithosphere is formed nor old lithosphere is destroyed. Transform plate boundaries are associated with transform faults. They are (like strike-slip faults) developed about vertically to the Earth's surface, and are created as a result of balance movements, due to the spherical form of the plate surface and the geometry of plate motion (Figure 3). California's San Andreas Fault is a transform boundary between the North American plate and the Pacific plate, which joins the east Pacific mid-ocean ridge to the subduction zone below the Cascade Range. Transform faults also cut up laterally all of the mid-ocean ridges into smaller fragments (Figure 3).

3. Driving Mechanism

Why do plates move? It is still not very clear what forces cause plates and continents to move around Earth's surface.

The most important motor of plate movement is mantle convection. Mantle convection is the process that continuously stirs the entire mantle as old, cold plates sink at subduction zones to the core mantle boundary, and hot rocks rise at the same time towards Earth's surface beneath a spreading center, forming new lithosphere to replace that lost to subduction (see *Mantle and Core*).

Although Earth's mantle is mostly solid rock, it is so hot that over geologic time it flows slowly at rates of approximately 10–20 cm/a. We simply assume that the mantle attempts to offset through convection the different temperature between core and surface. Recent results support the hypothesis that the entire mantle convects from the hot core–mantle boundary to the lithosphere. Heat from the core, supplemented by additional heat generated by radioactivity within the mantle, drives the entire mantle and lithosphere in huge cells of convecting rocks. A tectonic plate is the upper portion of a convecting cell, and thus glides over the asthenosphere because of the convection (Figure 2). Through the mantle convection, the upward branch of the convection transports great quantities of heat from the Earth's interior to the surface, where the transported hot rocks are cooled and a quantity of the heat radiates into space. In addition, cool, dense lithosphere sinks at the subduction zone to the Earth's interior. So, the Earth would have been much hotter at the time of its creation than it is today and the lithosphere was probably warmer and lighter in order to sink. Furthermore, the heat production of Earth is slowly decreased, as the radioactive isotopes continue to decay and they become less abundant. Therefore, we could assume that in a few billion years from now the Earth will be cool and the mantle strong. Consequently, the principles of plate tectonics and continents movement are only decisive for the form of Earth today.

Simultaneously with the mantle convection, some other processes and forces may contribute to the movement status quo of the lithospheric plates: The base of the lithosphere slopes downward from a spreading center (Figure 2). From this side, gravity causes the lithosphere to slide away from the spreading center over the soft, plastic asthenosphere (“ridge push”). In addition, the lithosphere becomes denser, thicker, and cooler as it moves away from a spreading center. The old lithosphere, far from the spreading center, becomes denser than the asthenosphere below. As a result, it can no longer float on the asthenosphere, and sinks deep into the mantle in a subduction zone, dragging the trailing portion of the lithosphere over the asthenosphere (“slab pull”). The frictional forces at the transform boundary between lithosphere and asthenosphere also play their role on the movement regime of the lithosphere.

If segments of continental crust or a great volcanic complex are led into a subduction zone, because of their low density, they cannot sink deep into the mantle so simply. As such obstacles come into a subduction zone, the result can be modifications to the movement direction or convergence rate of the plates concerned, until the convergence processes are complete.

Parallel to the huge curtain-shaped mass of mantle that rises beneath a spreading center, another branch of upward flowing mantle exists, called mantle plume. In contrast to mantle convection, mantle plume is a relatively small column of hot, plastic mantle rock rising vertically (Figures 2 and 3) (see *Mantle and Core*).

Plumes commonly rise from great depths in the mantle, because the rocks near the core-mantle boundary are hotter and more buoyant than the surrounding regions of the mantle. Others may form as a result of local heating in shallower portions of the mantle. As pressure decreases in the rising plume, magma forms, and rises erupting from volcanoes at a hot spot on Earth's surface. The Hawaiian island chain is an example of a volcanic center at a hot spot away from a plate boundary (see *Volcanology: Volcanic Activities, Chemistry and Effects on Environment*).

4. Vertical Movement of the Lithosphere

Besides the horizontal movements of Earth's lithospheric plates, vertical movements also affect the lithosphere. Lithospheric vertical movements are related to the overload or mass removal from the lithosphere. If a large mass is added to the lithosphere, it settles, and the underlying asthenosphere flows laterally away from that region to make space for the settling lithosphere. In contrast, if a large mass is subtracted from the lithosphere it emerges and the underlying asthenosphere ascends.

The concept that the lithosphere is in floating equilibrium on the asthenosphere is called Isostasy and the vertical movement in response to a changing burden is called isostatic adjustment. For example, when a glacier grows, the weight of ice forces the lithosphere downward. Conversely, when the glacier melts, the continent uplifts, it rebounds.

The lithosphere undergoes the same isostatic movements when a region is denudated and the eroded material is deposited in another adjacent region. The eroded region uplifts and the other settles. Consequently a thick basin can be formed. Furthermore, gravitational collapse or extensional collapse of an over-thickened crust is associated with crustal unroofing caused by isostatic rebound of the thickened crust (Figure 4).

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Biographical Sketch

Adamantios Kiliias has been professor at the Department of Geology and Paleontology of the Aristotle University of Thessaloniki since October 15, 1997. Prof. Kiliias was born on June 11, 1951 in Thessaloniki, Greece, and is of Greek nationality. From 1977 to 1981 he elaborated his dissertation at the University of Thessaloniki (Greece) with a theme about the geology of the north Pelagonian zone in Greece. From 1975 to 1977 he did postgraduate studies at the University of Vienna, and visited the University of Tuebingen in Germany as guest professor in 1988. He worked as scientific co-operator at the Greek Institute of Geological and Mineral Researches (IGME) (1981), at the Department of Geology of the National Metropolitan Polytechnic School of Athens (EMP) (1981), and at the Department of Geology and Paleontology of the University of Thessaloniki (1982). He was at the Department of Geology and Paleontology of the University of Thessaloniki, lecturer from 1982 to 1986, assistant professor from 1986 to 1990, and associated professor from 1990 to 1997. In October 1997 he was appointed professor at the same department and university. Prof. Kiliias received in December 1995 the "highest prize" of the Athens Academy for research about the geology of northern Greece. He was convener in the international congress of the European Association of Geophysics at Nice (France) in 1977, and member of the Scientific Committees of the Geological Greek-Bulgarian Symposium in Thessaloniki in 1989. His research field is structural geology. He has studied mainly the structural evolution of the Hellenides and their extension in neighboring regions in Albania, Bulgaria, Skopje, and Turkey. He has also worked in the Betic Cordillera, south Spain. The results of the investigations were published in international and Greek journals and are contained in about 70 original scientific papers. Many of the papers were also presented at one of over 30 international congresses.