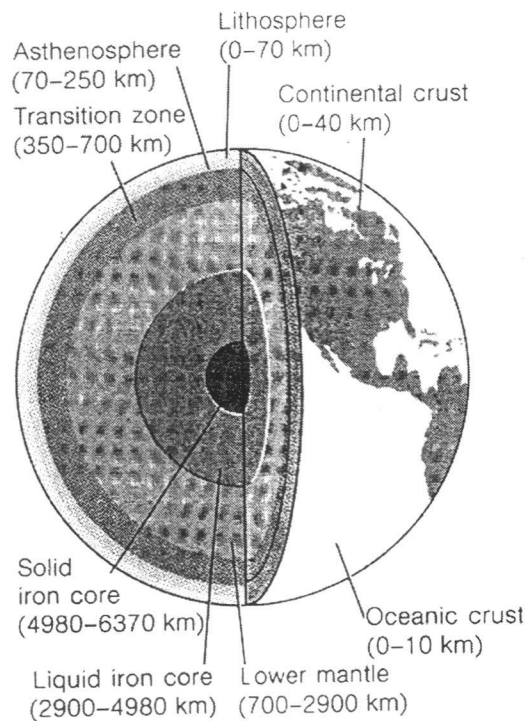


# PLATE TECTONICS AND EARTHQUAKES

K.V. Wilbert Kehelpannala

## Interior of the Earth

The Earth is a dynamic planet and formed about 4600 million years ago, together with the Sun and the other planets in the Solar System. Scientists believe that growing planets were molten - at least partly, and at least once. At such times, their component materials had the opportunity to separate according to melting points and densities. After the accretion of the Earth, the temperature of its interior increased due to many reasons. After 1000 million years the interior heated to the melting point of iron at depths between 400 km and 800 km, and iron began to melt in this region. The heavy materials, especially iron, sank towards the interior of the Earth, creating what is called the **core**, and light ones rose to the surface, creating the outer layer of the Earth, called the **crust** (Figure 1). The region between those two is called the **mantle**, which itself became layered according to chemistry and density. The process of gravitational separation is usually called differentiation, which was responsible for the formation of a layered Earth, as shown in Figure 1. This differentiation has also created a light crust depleted in iron and enriched in O<sub>2</sub>, Si, Al, Ca, K and Na.



**Figure 1.** Interior of the Earth and its internal layers. The early Earth was probably a homogeneous mixture with no continents or oceans. In the process of differentiation, iron sank to the centre and light material floated upwards to form a crust. As a result, the Earth became a layered planet with a dense iron **core**, a surficial **crust** of light rocks, and between them a residual **mantle**.

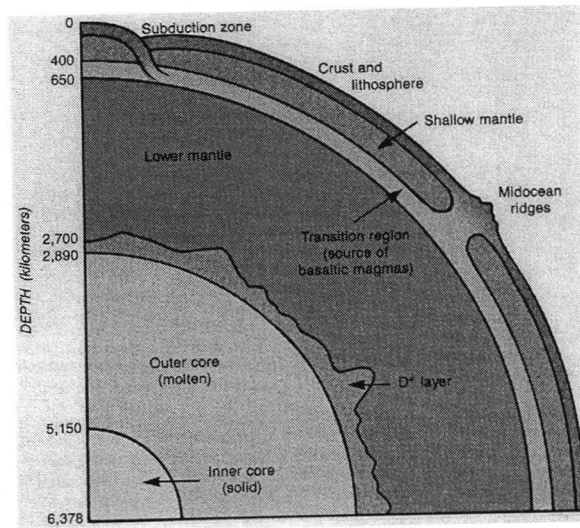


Figure 2. Cross-section of the interior of the Earth showing different layers.

Earthquakes that occur in tectonically active regions of the world generally generate a large amount of elastic energy, which transmits through the Earth as seismic waves. The wave fronts of earthquakes are refracted (bent) and reflected by discontinuities and gradients in material properties in the Earth's interior and recorded at seismic stations on the surface of the Earth. The seismic waves are of two types (see later). These are (i) body waves and (ii) surface waves. Velocities of these seismic waves depend upon elastic moduli and density of the medium through which they travel. From analysis of seismic waves, seismologists have confirmed the layered nature of the Earth's interior and have divided the planet into a crust, mantle and core and into numerous smaller subdivisions such as the upper mantle, transition region, lower mantle, outer core and inner core (Figure 2; Table 1).

Table 1. Summary of the Earth's internal structure.

Region	Depth (km)	Percent of Earth's mass	Percent of mantle-crust mass
Continental crust	0-50	0.374	0.554
Oceanic crust	0-10	0.099	0.147
Upper mantle	10-400	10.3	15.3
Transition region	400-650	7.5	11.1
Lower mantle	650-2,890	49.2	72.9
Outer core	2,890-5,150	30.8	
Inner core	5,150-6,370	1.7	

### Crust

The crust consists of the region above the Moho and is divided into two as oceanic crust (containing oceanic part) and continental crust (continental part). The thickness of the crust changes from about 3 km in some oceanic regions to about 80 km in some continental areas.

## Upper mantle

The upper mantle extends from the top (~ 650 km depth) of the lower mantle to the base of the crust (Figure 2). The upper mantle and the crust are separated by a seismic discontinuity called *Mohorovicic* discontinuity or *Moho*. The depth to the Moho varies from place to place: 10-12 km beneath the oceans and 30-35 km beneath the continents. The rocks of the base of the oceans (oceanic crust) are derived from the upper mantle.

## Lower mantle

The lower mantle is the largest subdivision of the Earth. It extends from the top of D" layer (about 2740 km depth) to the major seismic discontinuity at a depth near 650 km (Figure 2).

## Outer core

This extends from the boundary of the inner core (5155 km depth) to the core-mantle boundary, which is at a depth near 2900 km (Figure 2). The outer core, about 30% of the Earth's mass, is liquid and composed of iron-sulphur mixture containing roughly 12% sulphur and probably 2% nickel.

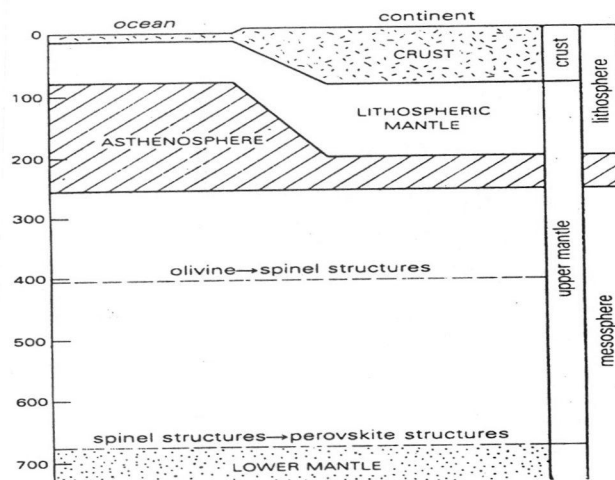


Figure 3. The major regions of the upper 700 km of the Earth.

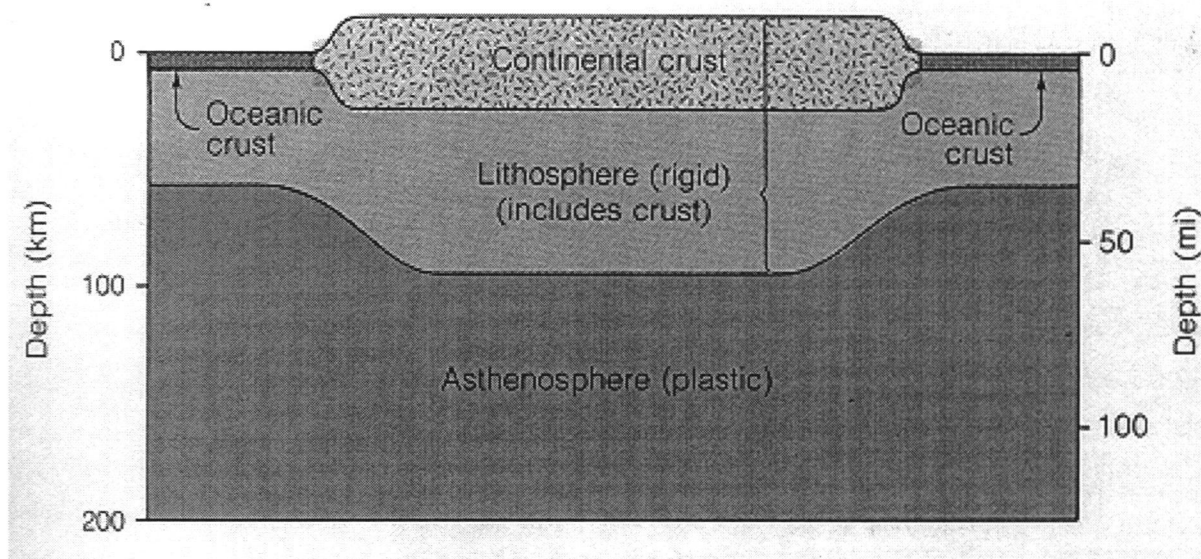
## Inner core

The inner core, 1.7% of the Earth's mass, is solid and composed of iron-nickel alloy (probably about 20% Ni and 80% Fe). The inner core is about the size of the moon.

## Lithosphere and Asthenosphere

According to the seismic velocity changes, the upper 700 km of the Earth has been subdivided into three main regions (Figure 3). These are, from top, lithosphere, asthenosphere and mesosphere (Figures 1, 3). The lithosphere contains the complete crust and the upper part of the upper mantle and its thickness varies from 50 km to 150-200 km (Figures 3, 4). This is the outer layer of the Earth (Figure 4). The lithosphere is strong and rigid and can easily be broken into large pieces called plates, which move about the surface according to the plate tectonic theory, which is one of the topics of this course. The

asthenosphere is a weak layer, on which the lithospheric plates move, and extends from the base of the lithosphere down to about 250 km (Figures 1, 3).



**Figure 4.** Figure showing the lithosphere and the asthenosphere of the Earth. The Earth's outer layer is called the lithosphere. It is made of the rigid upper mantle and the crust. The lithosphere moves on the asthenosphere, part of the mantle that flows.

### The Earth's magnetic field

Some minerals (e.g. magnetite) and rocks (e.g. ironstone and other iron-rich rocks) in the Earth's crust exhibit magnetic properties. These materials have acquired their magnetism from the Earth's magnetic field which existed at the time of their formation. Not all the planets have a magnetic field associated with them. For example, Venus and Mars show no magnetic fields. To a good approximation, the magnetic field at the surface of the Earth is like that due to a powerful bar magnet situated at the Earth's centre and aligned roughly along the axis of its rotation. Since the core of the Earth is likely to be made largely of iron, it might be thought that the solid inner core behaves like a permanent magnet. Scientists have found that the strength and position of the Earth's magnetic field have changed with time. The Earth's magnetic field is a highly variable and dynamic system on all scales of time, a system that has been in existence for most of its history and which is not due to permanent magnetism in the Earth's core. How, then, is the magnetic field generated in the Earth's core? The Earth's magnetic field is thought to originate as a self-exciting dynamo. The Earth's outer core, which consists of fluid iron, generates its own magnetic field. Large-scale fluid movements in the core seem to offer the best possibility for explaining the Earth's magnetic field. The magnetic field of the Earth has reversed many times in the past, which is referred to as **geomagnetic reversal**. The reversal of the Earth's magnetic field has confirmed the existence of plate tectonics (movements of lithospheric plates).

### Plate Tectonics

The lithosphere of the Earth is fragmented into many parts, of which about 10 are very important. These fragments, which move over the asthenosphere of the Earth, are called "plates". **Plate tectonics** is a theory that the outer shell (lithosphere) of the earth is made up

of thin, rigid plates that move relative to each other. The theory of plate tectonics was formulated during the early 1960s, and it revolutionized the field of geology. Scientists have successfully used it to explain many geological events, such as earthquakes and volcanic eruptions as well as mountain building processes and the formation of the oceans and continents. Major parts of a lithospheric plate are shown in Figure 5, and Figure 6 shows the major lithospheric plates of the Earth and their boundaries.

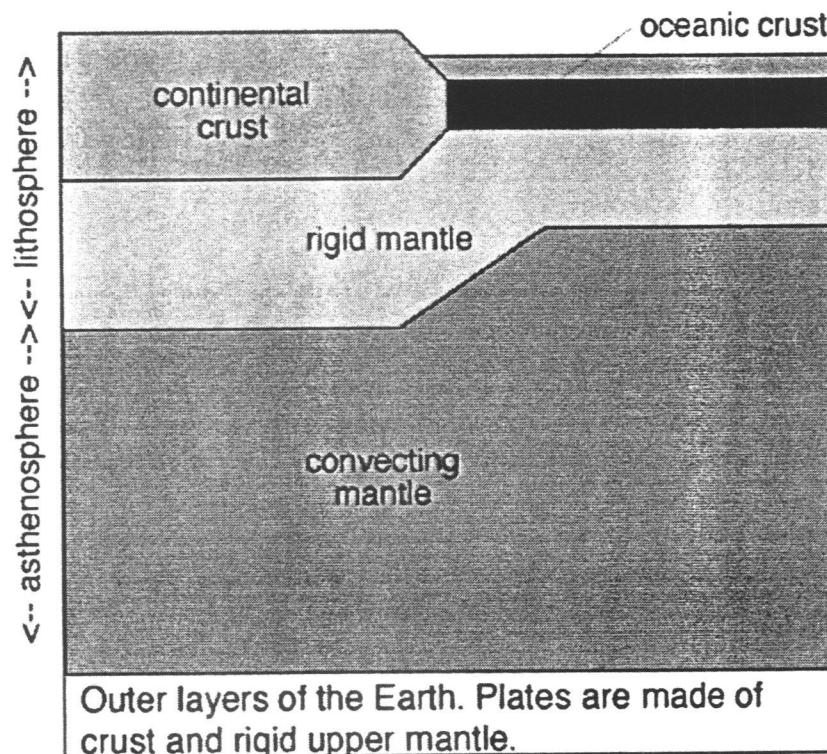


Figure 5. Schematic diagram showing major parts of a lithospheric plate.

Generally lithospheric plates are of two types. They are: (1) continental plates and (ii) oceanic plates. The continental plates consist of continents, together with some parts of the oceanic crust (Figure 5). The North American, South American, Eurasian, African, India-Australian, Antarctic and Arabian plates are examples for continental plates (Figure 6). The Pacific, Nazca and Phillipine plates are present-day oceanic plates (Figure 6). Plates are separated by plate boundaries that are of three types (Figures 7, 8, 9). These are

- (i) **divergent plate boundaries** (mid oceanic ridges and rift zones) (Figure 7),
- (ii) **convergent plate boundaries** (subduction zones) (Figures 8, 9, 10) and
- (iii) **transform faults** (Figure 11).

At divergent plate boundaries, plates move apart creating a new sea floor between the moving plates (Figure 7a). The Atlantic Ocean (Figure 7) had opened in this manner about 180 million years ago, when the South American plate separated from the African Plate. At the initial stages of opening of a sea at divergent plate boundaries, rift zones are formed (Figure 7a). Because new crust forming materials (basaltic rocks) are added to the oceanic crust at divergent plate boundaries, these are also known as constructive plate boundaries.

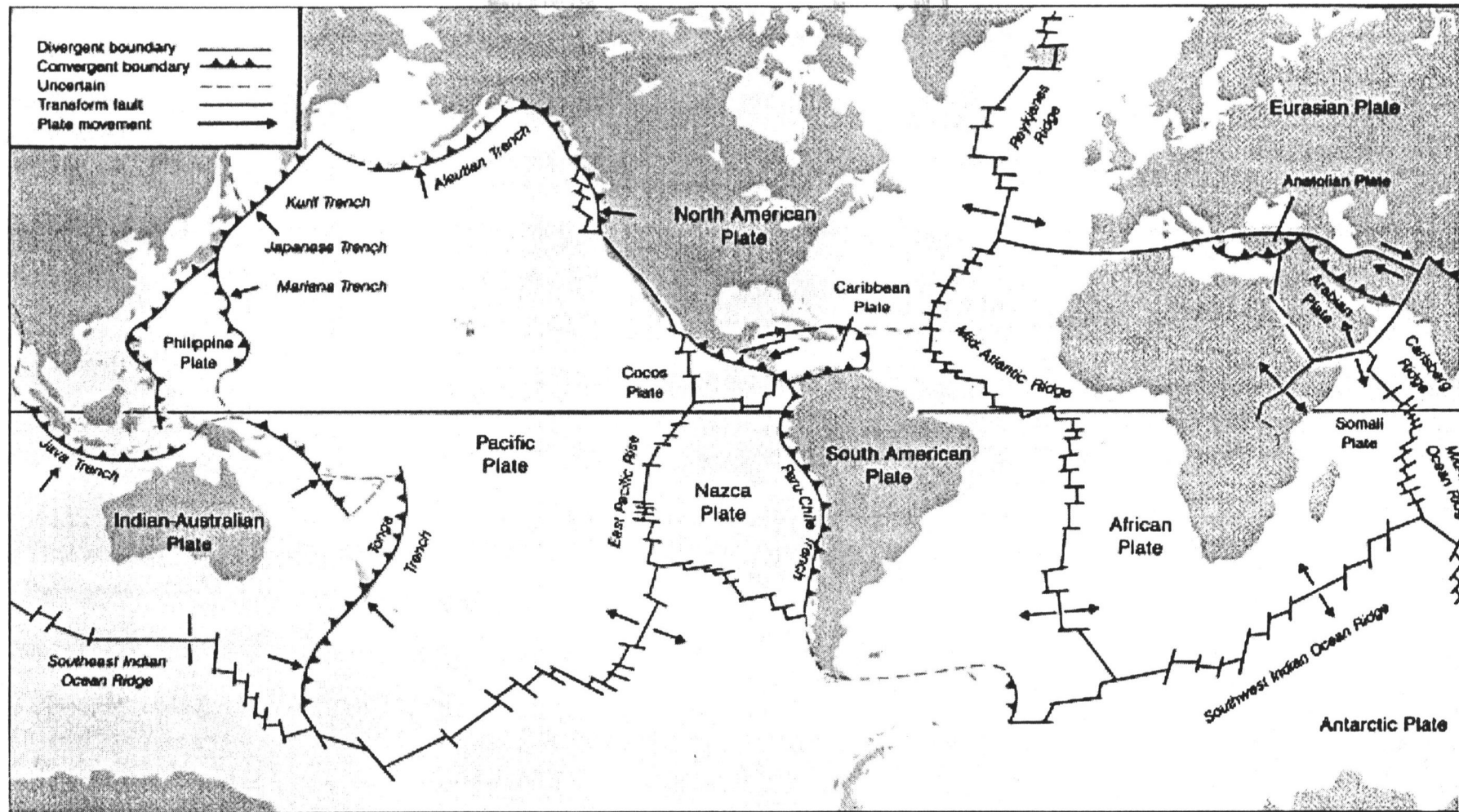


Figure 6. Major lithospheric plates of the Earth and their boundaries. At divergent plate boundaries (mid oceanic ridges) the plates move apart, only to collide and overlap at convergent boundaries (subduction zones). Plates slide past each other along transform faults.

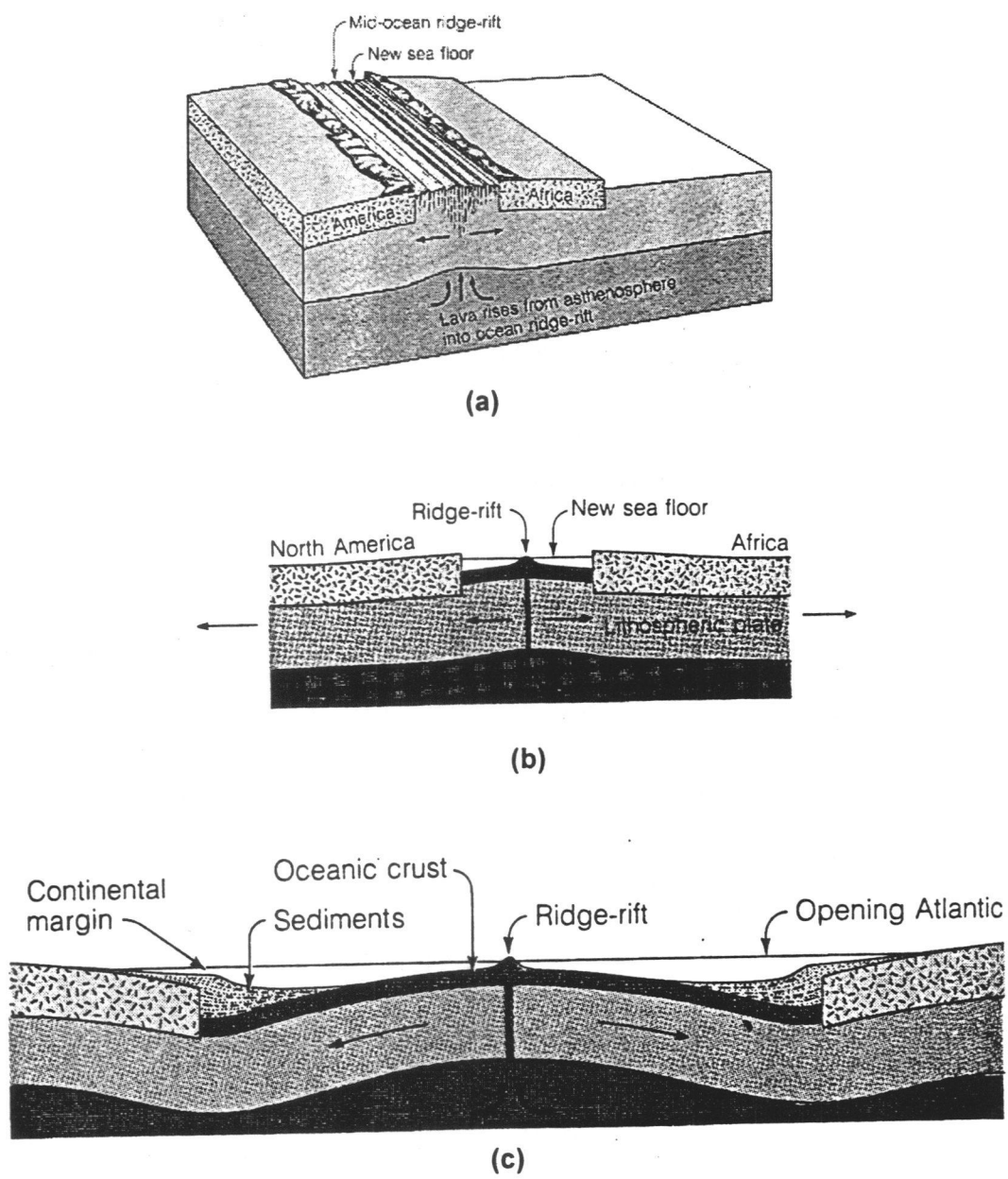
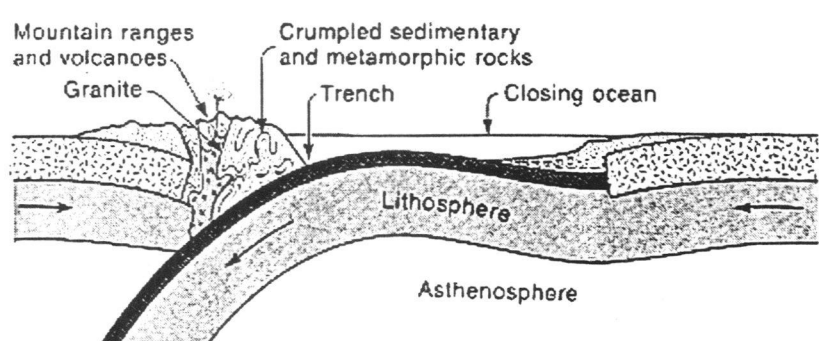
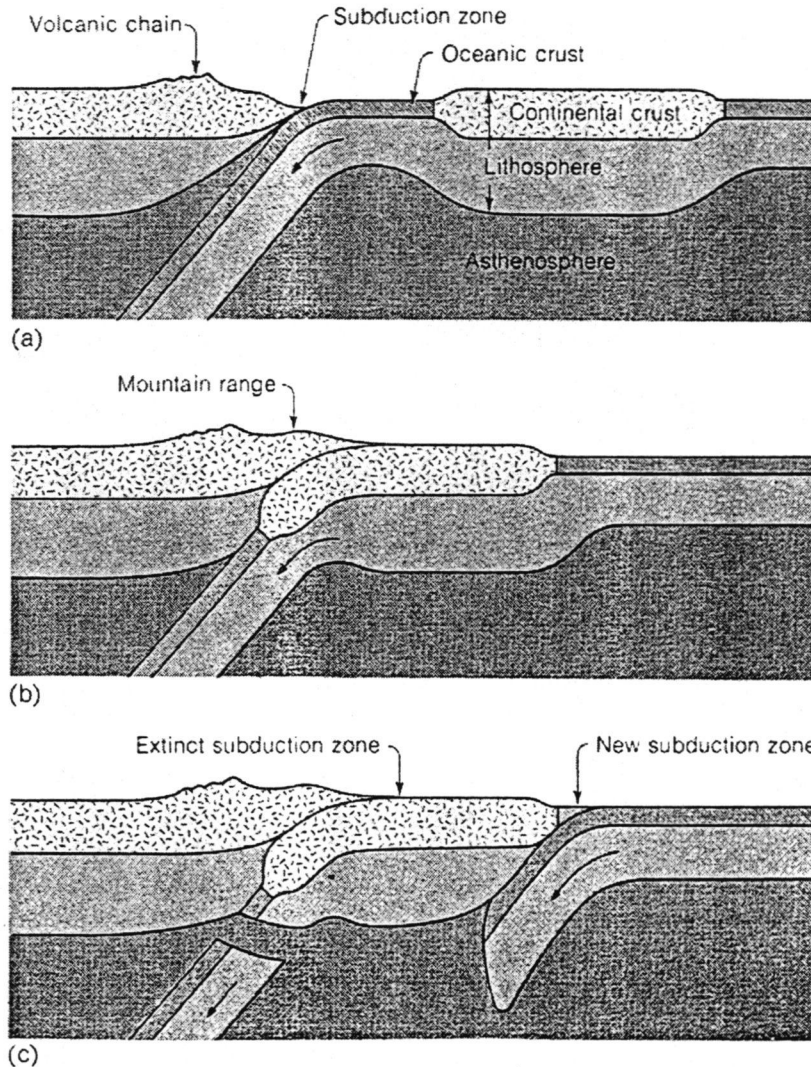


Figure 7. Cartoon showing a divergent plate boundary (mid-ocean ridge or rift zone) where two plates are moving apart, creating a new ocean or widening existing ocean.



**Figure 8.** A convergent plate boundary or a subduction zone, where two plates are moving towards each other. The sediments deposited at the margin of the overlying plates are crumpled, deformed and metamorphosed. Volcanos can also form at these boundaries.

In contrast, at convergent plate boundaries (Figure 8), plates move towards each other, and finally they collide to form mountain belts (Figures 8, 9, 10, 11). The collision can occur between continental-continental plates, continental-oceanic plates and oceanic-oceanic plates (Figure 11).



**Figure 9.** Possible stages of plate collision. (a) Convergence between plates with continental and oceanic lithosphere. Volcanic belts, mountains and earthquake belts are features of the overriding continental boundary. (b) Collision of continents, producing a mountain range and a volcanic belt. (c) Alternatively, the plate may break off and a new subduction zone be started elsewhere.

At the beginning of a continental collision, the oceanic crust of the involving plates may collide first (Figures 8, 9a). Subsequently, the continental parts may collide into a single landmass, closing the ocean between them (Figure 9b). The collisional zones are characterized by geological features called subduction zones, where the heavier plates subduct into the asthenosphere under the overlying plates (Figures 8, 9a, 11). The oceanic crust at subduction zones is deflected to form trenches that are the deepest parts of the oceans (Figure 8). At subduction zones or convergent plate boundaries, plates are subjected to destruction, and hence these boundaries are called destructive plate boundaries.



Figure 10. Part of the Andes mountain belt formed by the collision between the East Pacific plate and the South American plate. The rocks are here folded, faulted and uplifted.

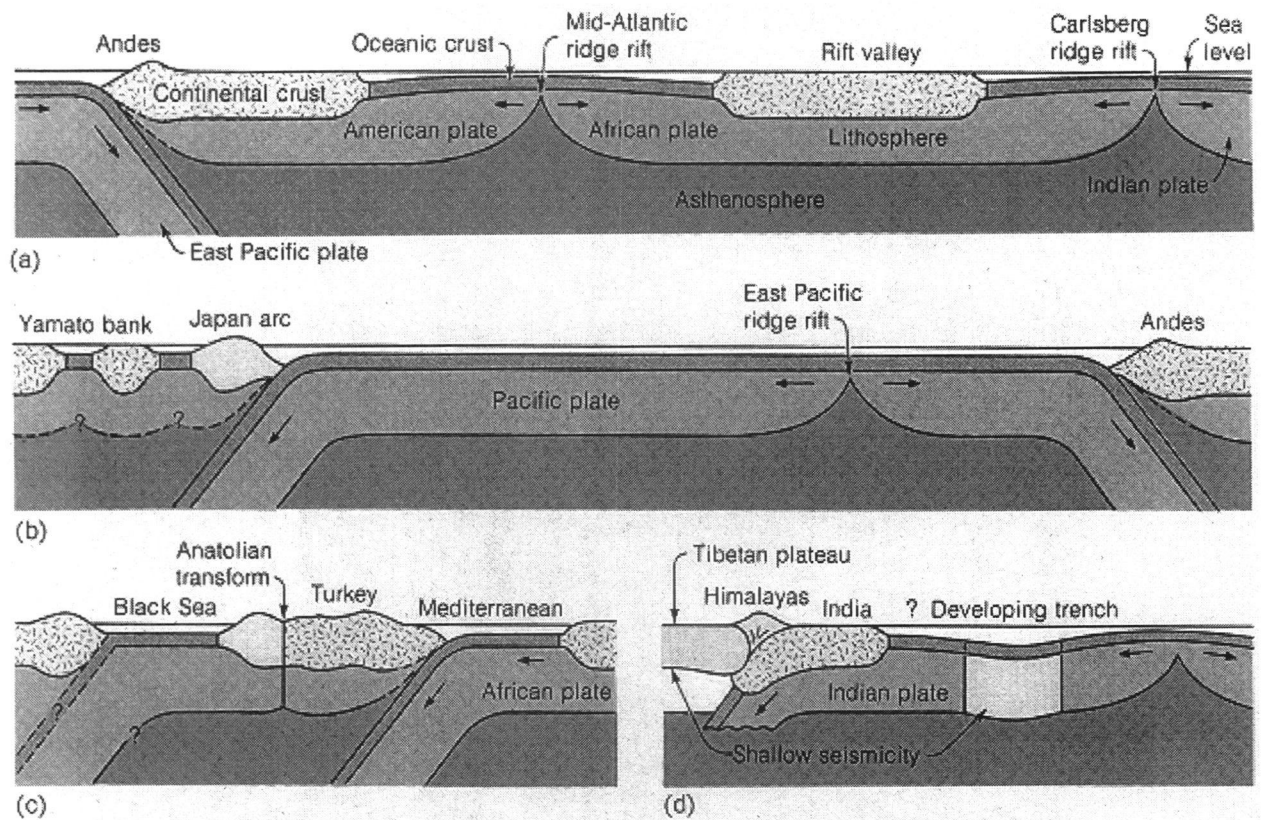


Figure 11. Schematic sections showing modern plate, ocean continent, and island-arc relationships.

All the mountain belts of the world have formed through collision between plates at subduction zones (Figures 6, 11). The mountain belts can be grouped into two classes. These are (i) interior mountain belts and (ii) peripheral mountain belts. Interior mountain belts form through the collision between two continental plates. The best examples for interior mountain belts are the Himalayas (Figures 6, 11d) and the European Alps (Figure 1). The Himalayas formed about 40 million years ago when the India-Australian Plate collided with the Tibetan Plate (or Eurasian Plate). This collision is still active and is responsible for earthquakes in India. Peripheral mountain belts form through the collision between oceanic plates and continental plates. The Rocky and the Andes in western America (Figures 6, 11a) are excellent examples for this type of mountain belts.

The third type of plate boundaries is transform faults along which plates slide past each other (Figure 12). There are many transform faults in the Earth (Figure 6), and they are responsible for changing the configuration of the mid oceanic ridges (Figures 6, 12). The San Andrea's fault in California and the fault, which separates the North American plate from the Caribbean plate, are two major transform faults (Figure 6).

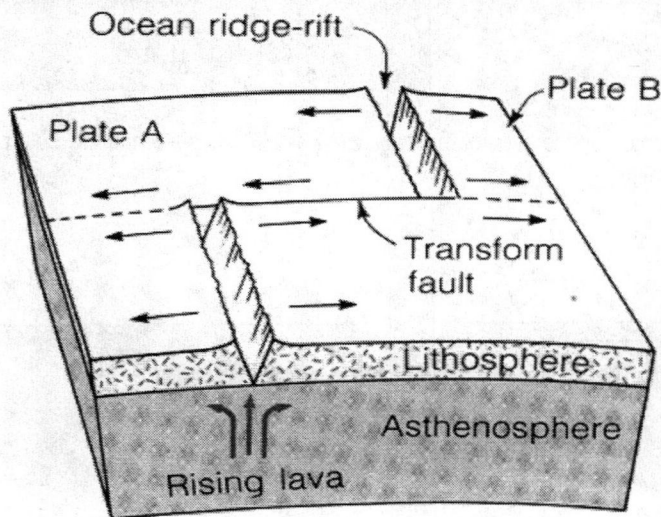
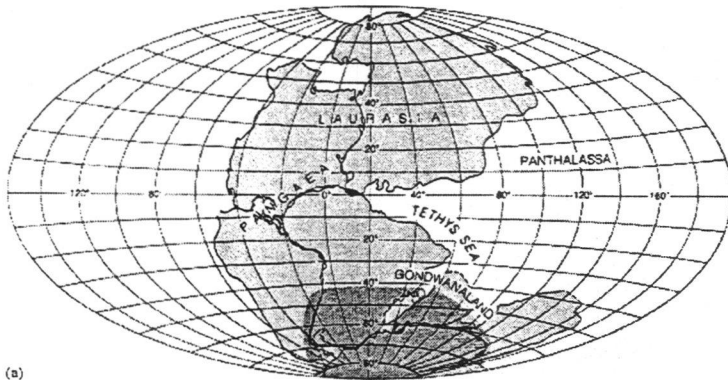


Figure 12. Schematic diagram showing a transform fault along which two plates (A and B) slide past each other.

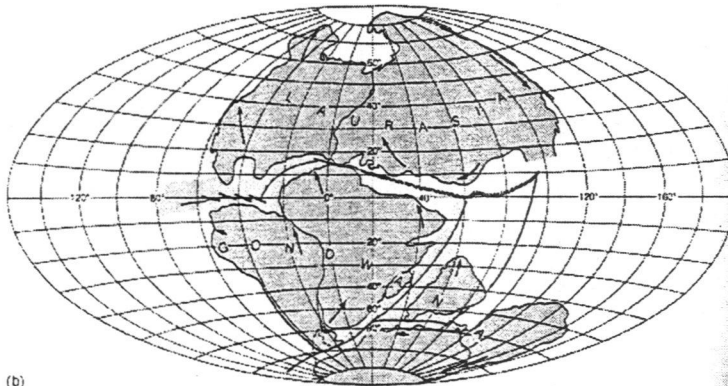
The break-up and collision of different plates accounted for the distribution of the present-day continents, which formed from a single supercontinent called "Pangaea" that existed about 200 million years ago (Figure 13a). The break-up of the Pangaea began about 180 million years ago with the separation of the Gondwanaland (landmass consisting of South America, India, Australia, Antarctica, Africa, Madagascar and Sri Lanka) from the Laurasia (continent consisting of North America and Eurasia) (Figure 5). Almost all continents were separated through rifting between 60 and 180 million years ago. Subsequent divergence and collision shaped the present-day configuration of the continents (Figure 13e).

Break-up of continents, collision through subduction and faulting has also existed before the birth of Pangaea. The supercontinent Pangaea had formed through collision and amalgamation of continents about 360 million years ago. The formation of supercontinents has also occurred during Archaean (>2400 million years ago) and Proterozoic (570 to 2300 million years ago) times. The formation and fragmentation of supercontinents seem to be a cyclic process, which is referred to as the "Wilson Cycle". Recently, geologists have established that the lifetime of a supercontinent cycle is about 500 million years. Sri Lanka is also a fragment of the Gondwanaland and may have formed through the collision of three different geological units (Figure 14) with different ages, about 600-500 million years ago.

also a fragment of the Gondwanaland and may have formed through the collision of three different geological units (Figure 14) with different ages, about 600-500 million years ago.  
**Figure 13. The break-up of supercontinent Pangaea**



(a) Ancient landmass Pangaea may have looked like this some 200 million years ago. **Panthalassa** evolved into the present Pacific Ocean, and the present Mediterranean Sea is remnant of the **Tethys Sea**. Permian (280-22 million years ago) glacial deposits are found in widely separated areas, such as South America, Africa, India, and Australia. This distribution is simply explained by postulating a single continental glacier flowing over the southern polar regions of Gondwanaland in Permian time, before the break-up of the continents. The dark shading shows the probable extent of this glacier.



(b) One view of world geography at the end of the Triassic period, about 180 million years ago, after some 20 million years of drift. Light grey colour shows the new ocean floor between separated continents. Arrows indicate motions of continents since drift began.



(c) Continents and oceans at the end of the Jurassic period, 135 million years ago, after some 65 million years of drift.



(d) Continents and oceans at the end of the Cretaceous period (65 million years ago). Light grey shows new ocean floor created after some 135 million years of drift.



(e) Present-day continents and oceans

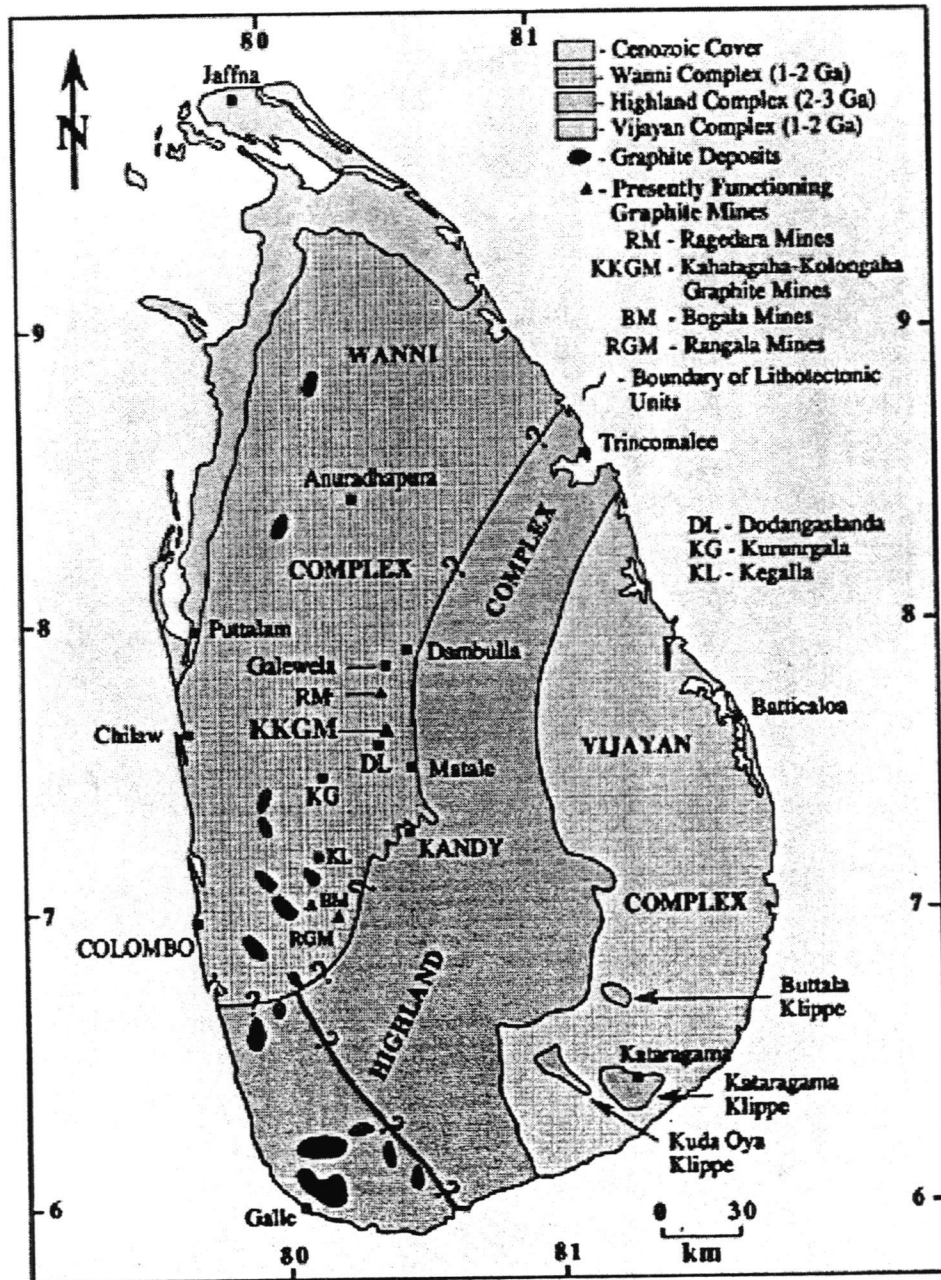
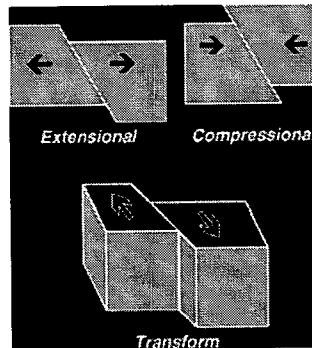


Figure 14. Map of Sri Lanka showing geological subdivisions. The central unit, called the Highland Complex, is about 2000 million years old, and the other two units, called the Wannai Complex and the Vijayan Complex, are about 1000 million years old. These three units were brought into contact by plate tectonism some 550-650 millions years ago.

## Earthquakes

### What is an earthquake?

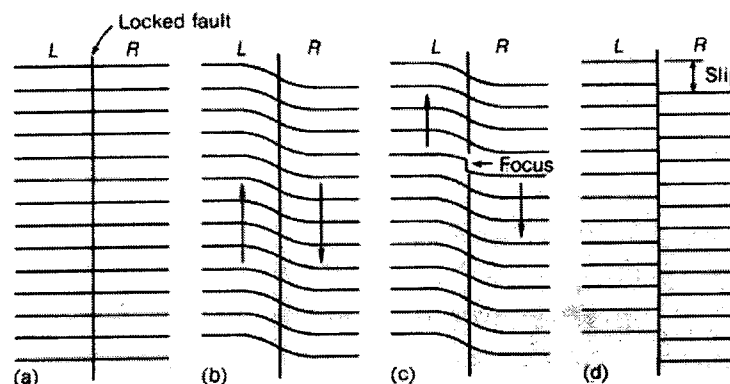
An earthquake is a sudden movement or trembling of the Earth caused by the abrupt release of slowly accumulated stress on a fault. A fault is a fracture plane along which rocks are displaced (Figure 15). An earthquake occurs as the result of a slow build-up of strain energy in rocks, usually caused by the relative motion along a fault. This strain energy is stored in rocks as elastic energy. When a fault or volume of rock can no longer resist movement, the stored elastic strain energy is suddenly released, causing an earthquake. The mechanism of an earthquake can be explained by a simple model called the **elastic rebound theory** (Figure 16).



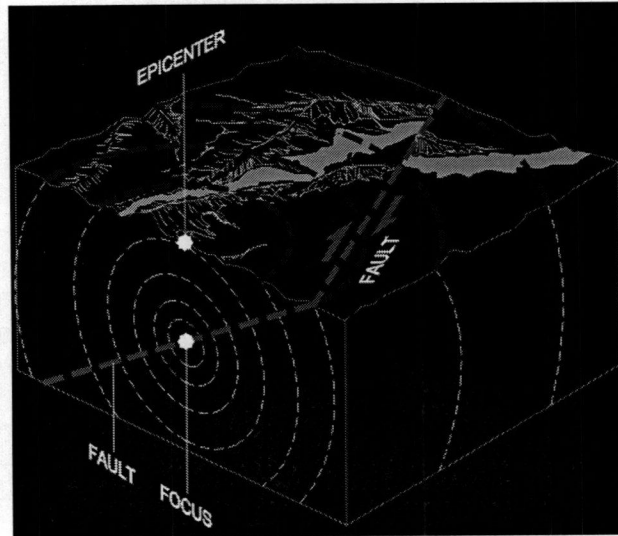
**Figure 15.** Different types of faults in the Earth. Earthquakes generally occur along faults. Faults form in extensional situations are called normal fault. Reverse fault form in compressional environments. In transform situation strike-slip fault are developed.

This model is diagrammatically shown in Figure 16. Over time, one side (L) of the fault is displaced relative to the other side (R). The deformation (change in form or shape) continues until the stresses (forces per unit area) on the fault are large enough to overcome the friction (frictional bond is broken) between the two blocks (L, R) of material; then an earthquake (sudden displacement or rupture) occurs (Figure 16d), and the strain is released as energy. Thus, the size of the earthquake is directly related to the friction in the fault.

The sudden displacement or rupture begins at a point, which is called the earthquake focus (Figure 16c, 17). The point on the Earth's surface immediately above the earthquake focus (location of the earthquake within the Earth) is called the earthquake epicentre (Figure 17). Once the rupture begins, it travels at a speed of about 3.5 kilometres per second (7200 miles per hour), continuing for as much as 100 kilometres. In large earthquakes, the slip, or displacement, of the two blocks (Figure 16 d) can be as large as 15 metres. Once the

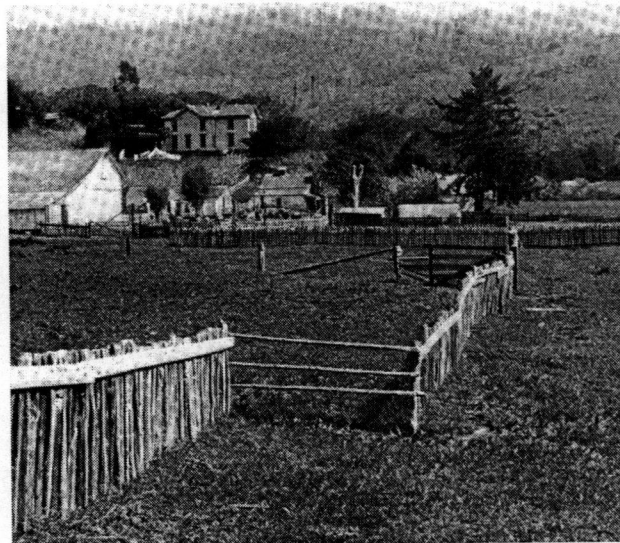


**Figure 16.** The elastic rebound theory of an earthquake. The two simulated crustal blocks, L and R, are being forced to slide past each other (a). Friction along the fault prevents slip (b), but the deformation builds up until the "frictional" lock is broken (c) and earthquake slip occurs (d).



**Figure 17.** The focus of an earthquake is the site of the initial slip on the fault. The epicentre is the point on the surface above the focus. Seismic waves radiate from the focus (dash circles).

frictional bond is broken, the elastic energy, which had been slowly accumulated over tens or hundreds of years, is suddenly released in the form of intense seismic vibrations (within a few minutes), which constitute the earthquake. These vibrational waves are propagated large distances in all directions from the fault. Destructions occur due to this vibrational energy and due to the displacement along the fault. An example of displacement caused by an earthquake is shown in Figure 18.

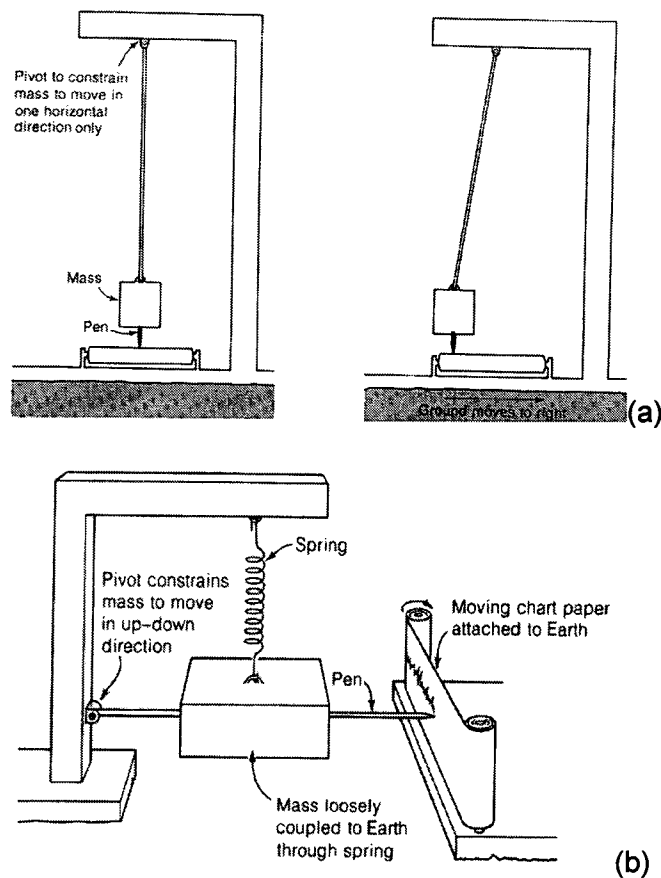


**Figure 18.** A fault formed during an earthquake that occurred in America. Note that the fence is displaced by the fault.

The elastic strain energy released during an earthquake propagates as seismic waves. Energy release gives the most precise measure of the size of an earthquake. The magnitude of an earthquake is given by the “**Richter magnitude scale**”, which is based on the amplitude of seismic waves recorded by instruments called seismographs (Figures 19, 20). Table 2 gives magnitude of earthquakes, energies, effects and other related statistics.

**Table 2. Magnitude, energies, effects and other related statistics of earthquakes.**

Characteristic Effects Of Shallow Shocks In Populated Areas	Approximate Magnitude	No. Earthquakes Per Year	Energy (Ergs)
Damage near total	$\geq 8.0$	0.1-0.2	$> 10^{25}$
Great damage	$\geq 7.4$	4	$\geq 0.4 \times 10^{24}$
Serious damages	7.0-7.3	15	$0.04-0.02 \times 10^{24}$
Considerable damage to buildings	6.2-6.9	100	$0.5-23 \times 10^{21}$
Slight damage to building	5.5-6.1	500	$1-27 \times 10^{19}$
Felt by all	4.9-5.4	1,400	$3.6-5.7 \times 10^{17}$
Felt by many	4.3-4.8	4,800	$1.3-27 \times 10^{16}$
Felt by some	3.5-4.2	30,000	$1.6-76 \times 10^{15}$
Not felt but recorded	2.0-3.4	800,000	$4 \times 10^{10}-9 \times 10^{13}$



**Figure 19. Principles of seismographs that record earthquakes. (a) A pendulum-mounted seismograph. (b) A spring-mounted seismograph to record vertical ground motions.**

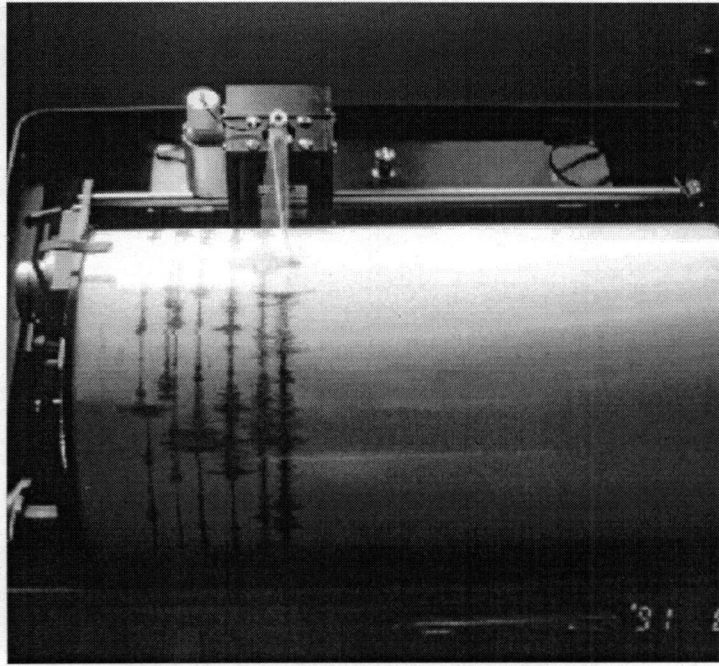


Figure 20. Seismograph recording of an earthquake.

In 1935 C.F. Richter, a seismologist, studied the local earthquakes in southern California and proposed a particular logarithmic magnitude scale for them. A number of other logarithmic magnitude scales for earthquakes have been proposed, all of which are based on measurements of the amplitude of the seismic waves. All the magnitude scales are of the following form:

$$M = \log_{10}\left(\frac{A}{T}\right) + q(\Delta, h) + a \quad (1)$$

where  $M$  is the magnitude,  $A$  the maximum amplitude of the wave (in  $10^{-6}$  metres),  $T$  the period of the wave (in seconds),  $q$  a function correcting for the decrease of amplitude of the wave with distance from the epicentre and focal depth,  $\Delta$  the angular distance from seismometer to epicentre,  $h$  the focal depth (depth to the focus) of the earthquake, and  $a$  an empirical constant.

Seismic waves are generally of two major types. These are body waves and surface waves. Body waves travel through the body of the Earth, and their propagation is similar to that of light. Body waves are reflected and transmitted at interfaces where the seismic velocity and/or density change, and they obey Snell's law. There are two types of body waves (Figures 21, 22): (1) P-waves or primary waves (compressional waves) (Figure 21) and (2) S-waves or secondary waves (shear waves) (Figure 22). P waves can travel through any medium (solid or liquid), whereas S-wave travels only through solid medium (S-wave does not propagate through liquids). P- and S-waves are very important in understanding the deep interior of the Earth.

Seismic waves are very important in the study of the interior of the Earth, which is illuminated by seismic rays radiating outward in all directions from the large number of earthquakes that occur in tectonically active regions (e.g. subduction zones, mid-oceanic ridges or rift zones and transform faults) of the world. The wave fronts generated at right angles to these ray paths are reflected and refracted by discontinuities and gradients in material properties in the interior of the Earth and are recorded at seismic stations on its surface. From analysis of such waves seismologists have divided the Earth into core, mantle and crust and into numerous smaller subdivisions such as inner core, outer core, upper mantle, transition region and lower mantle as mentioned earlier.

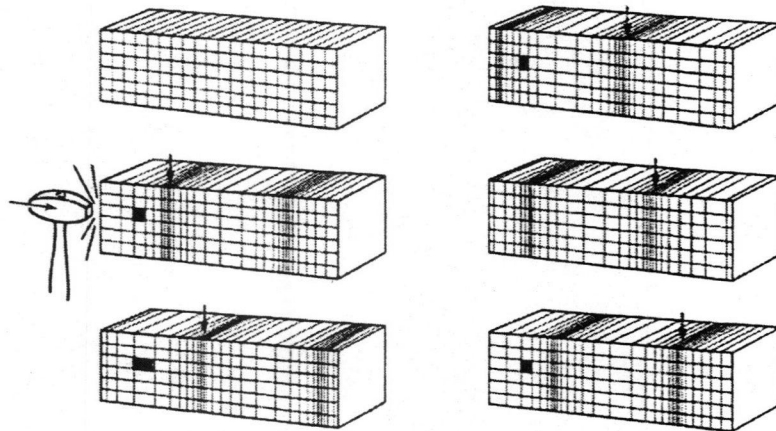


Figure 21. Stages of the deformation of a block of material with the passage of compressional P-waves.

P-wave and S-wave velocities depend on the physical properties (density and elastic moduli) of the material through which the waves travel (equations 2 and 3). The seismic velocities of P-wave and S-wave are calculated according to the following equations:

$$\text{seismic velocity of P-waves} = v_p = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}} \quad (2)$$

$$\text{seismic velocity of S-waves} = v_s = \sqrt{\frac{\mu}{\rho}} \quad (3)$$

where  $K$  is the bulk modulus or incompressibility,  $\mu$  the shear modulus or rigidity and  $\rho$  the density of the material.

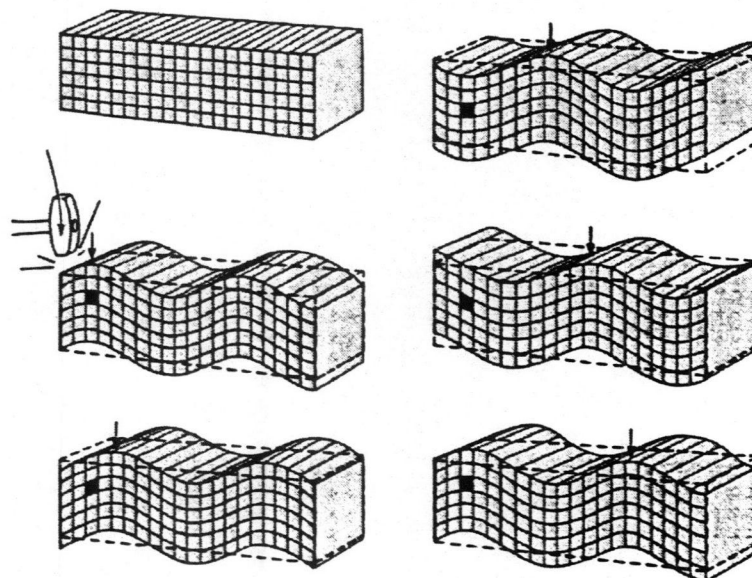


Figure 22. Stages in the deformation of a block of material with the passage of shear waves (S-waves).

The surface waves are seismic waves, which are guided along the surface of the Earth and the layers near the surface. They do not penetrate into the deep interior of the Earth. Surface waves are generated best by shallow earthquakes. Nuclear explosions do not generate comparable surface waves, and this important fact is the basis for one criterion of discrimination between earthquakes and nuclear explosions. There are two types of surface waves: (i) Love waves and (ii) Rayleigh waves.

### Where do earthquakes occur?

As already mentioned earlier, earthquakes are always associated with faults. Figure 15 illustrates different types of faults occurring in the Earth. Faults generally occur in association with major plate boundaries. This means that earthquakes should occur along all types of plate boundaries of the Earth. Almost all the epicentres of earthquakes occurring in the world follow plate boundaries (Figures 23, 24) such as rift zones or mid-oceanic ridges, subduction zones and transform faults (Figure 24).

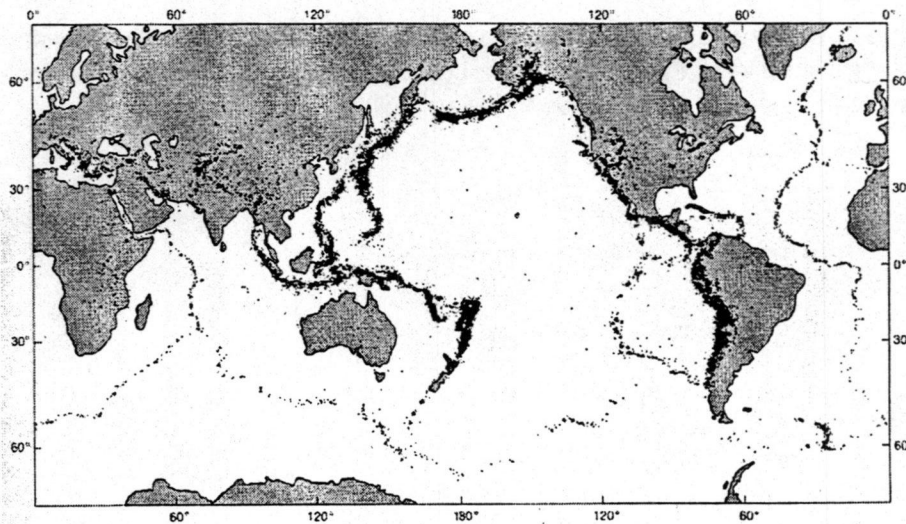


Figure 23. Epicentres of some earthquakes with focal depths between 0 and 700 km. All these earthquakes are located along plate boundaries.

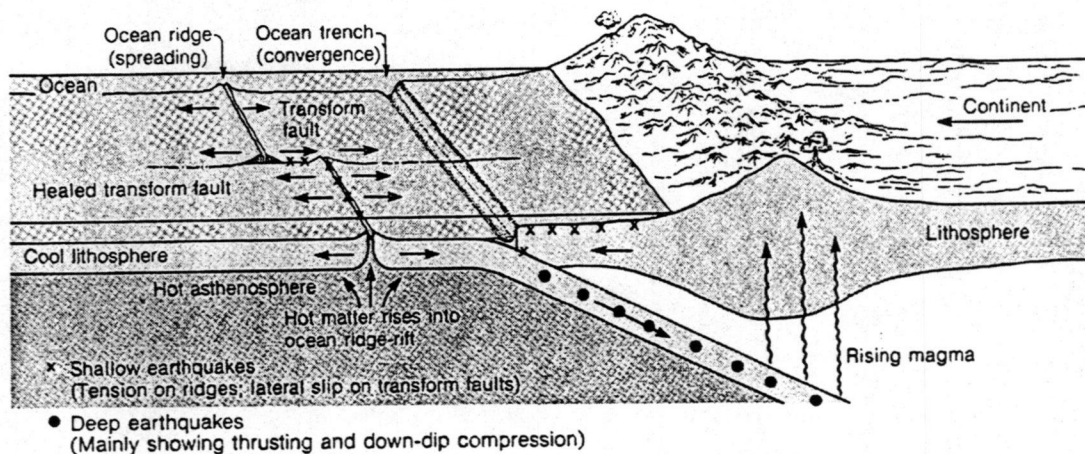


Figure 24. Schematic diagram showing the association of earthquakes with three types of plate boundaries: oceanic ridges (divergent plate boundaries), subduction zones (convergent plate boundaries and transform faults).

As shown in Figure 15, there are three major types of faults. These are (i) normal faults (in extensional environments), (ii) reverse faults (in compressional environments) and (iii) strike-slip faults (in transform situations). Normal faults are associated with rift zone and mid-oceanic ridges. Reverse faults originate at subduction zones, and strike-slip faulting occurs along transform faults. The association of earthquakes with three types of plate boundaries is shown in Figure 24.

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