CHAPTER 9: EARTH'S INTERIOR AND PLATE TECTONICS

In this chapter you will learn about what drives mountain building, causes regional metamorphism, generates magmas, and forms ocean basins, a system known as plate tectonics. For more information on plate tectonics and mountain building I suggest Park (2012).

(Note: Terms in red and italics appear as entries in the companion glossary.)

Now that we know something about Earth's crustal materials, placing past geologic events in order, and determining numerical ages for rocks, it is time to step back and see what drives this system. Earth has a whole fleet of processes that operate on a global scale. They move, deform, destroy, and create Earth materials. It is what generates magmas that form igneous rocks, triggers earthquakes, causes faults and folds, *uplifts* (or *subsides*) different areas of Earth's surface, makes mountain ranges, moves continents, triggers erosion, and can also influence Earth's climate. Think about how the planet would be different if these interior processes were not constantly triggering the changes that reshape Earth's surface. The surface would be a static place with no mountain ranges and dotted with craters, like the surface of the moon and some other planets, where erosion could not remove them. To understand how this system operates we first must have an appreciation for the structure and composition of Earth's interior.

9.1 EARTH'S INTERIOR AND HOW WE KNOW

Earth's interior is classified into different layers using two sets of criteria: 1) differences in chemical and mineral compositions, and 2) differences in how layers deform, i.e., whether they are brittle and strong, or ductile and soft. We refer to this second category as mechanical or *rheological properties*. We cannot drill a hole very far into Earth's surface to collect samples of the subsurface, so we must infer what is inside Earth from other evidence. In fact, the deepest hole ever drilled, about 12.2 km deep on the Kola Peninsula east of Finland in Russia, never even got deep enough to penetrate through Earth's outer layer, the crust. It didn't get close! (For more on Earth's deepest drill hole see: Kola Borehole.

There are several key observations that give us some information about Earth's internal makeup. Basaltic magmas generated at great depth will sometimes bring up inclusions that have minerals or rock types that were formed under very high pressures. However, these rocks are from relatively shallow parts of Earth's interior and never from very far below the crust. In a few places on Earth, deep rocks from about 10-15 km beneath Earth's surface have been thrust upward along faults in rock assemblages called *ophiolites*. These mafic rocks are exposures of ancient sections of the crust from beneath an ocean basin, but they are still not from very deep beneath the surface. From *seismology*, which is the study of shock waves that travel through Earth, we have been able to determine the depths of boundaries that either reflect, bend (refract), or attenuate (dampen or stop) earthquake waves where the density and mechanical properties of the rocks change. The types of waves being affected tell us something about the changes in rock properties that occur across these boundaries. We know from these studies that the internal layers are spherical, i.e., they parallel Earth's surface and have approximately the same depth everywhere, and there are major contrasts between the densities and mechanical properties of subsurface layers.

The physics of Earth's interior, or its *geophysical* properties, can provide insights into its subsurface structure. The strength of Earth's gravitational field is dependent on the mass of Earth, which allows us to infer the total mass of the Earth and places some constraints on Earth's mass distribution. It helps us calculate minimum densities for Earth's internal layers. Earth also has a strong magnetic field, which is generated in Earth's interior and places constraints on the types of materials that can exist in the subsurface. Finally, we get to examine meteorites that fell to Earth, which appear to be fragments of the interiors of planets that once existed in our solar system. They probably had interior layers with compositions like Earth's interior layers. These early planets broke apart during collisions with other planetary bodies or when bombarded by asteroids in the early history of the solar system. All our ideas about Earth's interior must be consistent with the observations above.

The center of Earth is very hot, and we tend to think of this heat as being left over from early in the history of the solar system, when the whole Earth started as a ball of magma. This molten stage lasted for a relatively brief time before cooling created a solid outer layer. The heat in Earth's interior today is not heat that is left over from billions of year ago, but instead is still being generated by radioactive decay, which releases heat that does not easily escape. Without the heat produced by radioactivity in Earth's interior the whole planet would have long ago cooled to a complete solid with a much lower temperature.

9.2 THE COMPOSITION OF EARTH'S INTERIOR

We will first look at the classification of Earth's interior layers according to their chemical and mineral compositions (Table 9.1 and Fig. 9.1). Beginning at the land surface and extending to a maximum depth of 60-70 km beneath the surface is Earth's *crust*. The crust is made of different rocks beneath the ocean basins as compared to beneath the continents. On the continents, the crust (*continental crust*) has an average composition of the igneous rock granite. Therefore, it is sometimes called *granitic* or *sialic crust*. Like granite, the continental crust is rich in quartz and feldspar. It has a relatively low density of about 2.7 g/cc and an average depth of 40 km. It can be thicker beneath large mountain chains, such as the Himalayas (up to ~70 km), or thinner where the crust is being pulled apart, as occurs in the Basin and Range Province of Nevada and eastern California (~20-30 km). Beneath the ocean basins is *oceanic crust*, which has a total thickness of about 10-15 km and is almost entirely made mafic igneous rock with a thin veneer of ocean sediment. For this reason, it is sometimes called *basaltic crust*. It has a higher density (3.0 g/cc) than the continental crust and is richer in iron and magnesium with less silicon and aluminum.

The bottom of the crust, a boundary known as the *Moho*, which is short for Mohorovičić Discontinuity, named after its Croatian discoverer, separates lighter rocks of the crust from denser rocks below. It represents a boundary where seismic waves change velocity because of a change in chemical properties that influence density. Beneath the Moho and extending to a depth of 2900 km, over 70 times thicker than the crust, is the *mantle*. The mantle is predominantly composed of iron and magnesium silicates and oxides with very little aluminum. As a result, it has a density of 4.5 g/cc, which is significantly higher than that of either continental or oceanic crust. Beneath the mantle is Earth's *core*. Contrary to science fiction movies and novels, we have no chance of drilling a hole to, or visiting, the core. The core is made of nearly pure metal. The outer core is liquid down to a depth of 5100 km, and the inner core is solid all the way to the center of Earth at 6370 km. Although it is very hot in the inner core, it is a

Table 9.1 *– Layers within Earth based on composition.*

CRUST

A. Continental (Granitic/Sialic) Crust
30-70 km, average thickness: 40 km
mostly Na, K, Ca, Al-silicates, and quartz
density: 2.7 g/cc

B. Oceanic (Basaltic) Crust
10-15 km
mostly Fe, Mg, Ca, Al-silicates and silicates
density: 3.0 g/cc

MANTLE
thickness is ~2900 km
composed of Fe, Mg oxides and silicates
density: 4.5 g/cc

CORE (liquid outer core; solid inner core)
depth to outer core: 2900 km
depth to inner core: 5100 km
composed of Fe-Ni alloy
electrically conductive
outer core generates the geomagnetic field
density: 10.7 g/cc, responsible for Earth's gravitational field.

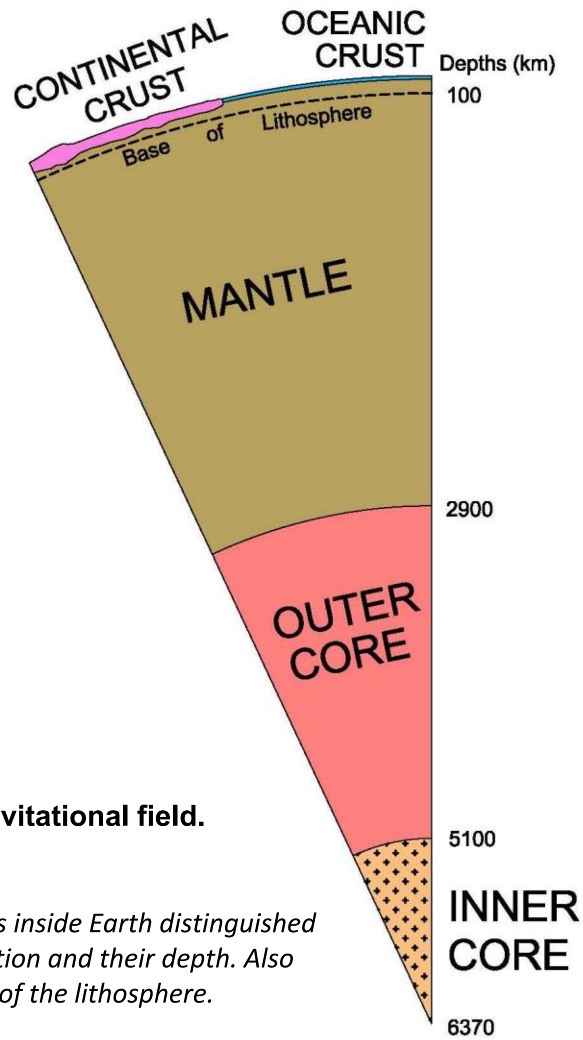

Figure 9.1 *– Layers inside Earth distinguished based on composition and their depth. Also shown is the base of the lithosphere.*

solid because of the extreme pressure at this depth. The core's mass and magnetic properties are best explained as a metallic alloy of iron and nickel. The metallic liquid in the outer core, which conducts electricity, is responsible for generating Earth's strong magnetic field. The high density of the whole core at 10.7 g/cc accounts for Earth's relatively high mass and strong gravitational field. Meteorites that crash to Earth and are made of solid iron-nickel metal alloy appear to be pieces of the core of an Earth-like planet that was blasted apart during collisions in the early history of the solar system. These metallic meteorites serve as an analog for the material in Earth's core. Metallic meteorites have a distinctive Widmanstätten crystallization structures. See Widmanstätten pattern.

9.3 MECHANICAL PROPERTIES OF EARTH'S INTERIOR

Earth's crust and mantle can also be subdivided into layers that have different mechanical properties and are independent of composition. The outer 100 km of Earth, which includes the whole crust and the uppermost mantle, is a layer called the *lithosphere* (Fig. 9.1). Rocks of the lithosphere are relatively rigid and brittle. Beneath the lithosphere rocks are still solid, but they are much weaker and ductile, and they behave more like silly putty. This weaker layer is called the *asthenosphere*, and it extends to the bottom of the mantle. In the asthenosphere, rock is not rigid, and as a result it responds to temperature differences that influence density. Warmer areas with lower density rise and cooler areas with higher density sink, creating a circulation pattern in the asthenosphere by a process known as *convection*. Hot, less dense materials rise relative to cool, more dense materials, and this transfers heat from Earth's interior out to the surface. This is what happens when hot air rises to the ceiling in a room. When it reaches the ceiling, it spreads out across the ceiling. It stays there until it cools, becomes denser, and sinks back to the floor. However, when convection occurs in the asthenosphere, the very sticky rock material of the asthenosphere drags the overlying rigid lithosphere with it. Separate areas of the lithosphere, known as *lithospheric plates*, are dragged in different directions and move relative to each other. Where plates sink into the asthenosphere, they can also exert a drag on the rest of the plate, which contributes to the plate's movement. The movement of lithospheric plates, and the resulting deformation we see at Earth's surface, is known as *plate tectonics*, which is responsible for slowly changing the positions of plates and continents over time.

9.4 PLATE TECTONICS

9.4.1 Continental Drift and Plate Tectonics

In 1912, *Alfred Wegener*, a German meteorologist and arctic researcher, proposed that the modern continents had once been a single large land mass that split apart. He observed that the outlines of the continents on opposite sides of the Atlantic Ocean seem to tightly fit together when placed next to each other. In support of this idea, he showed that the rock types, structures, and fossils on different continents matched each other where he thought the continents were once connected. This became known as the theory of *continental drift*. However, Wegener's theory was mostly rejected at the time. The main obstacle was the lack of a mechanism that could cause the continents to plow through the rigid rocks at Earth's surface. At the time, very little was known about the shape and structure of the ocean basins, or Earth's interior.

During World War II, there was a tremendous leap in technology, especially the development of sonar devices that for the first time allowed us to remotely map the deep ocean floor. The mid-ocean ridges in the centers of the ocean basins were explored, as were the deep ocean trenches along the perimeter of the Pacific Ocean. Development of the field of *paleomagnetism*, the study of the magnetism locked into rocks at the time they formed, allowed us to determine how the ocean floor and continents have moved through time. Seismologists also plotted the positions and depths of earthquakes, which were very numerous along certain boundaries and minimal in most other areas. The result of all this new information was the development of a comprehensive theory that explained not only the movement of the continents but also the lithospheric plates on which they rested. This is the theory of *plate tectonics*. Plate tectonic theory says that the lithospheric plates move relative to each other. At their boundaries, crust and lithosphere are either created or destroyed. The movement of the plates is due to convection in the aesthenosphere, which results in slow circulation of soft asthenospheric rock that drags the lithospheric plates around the globe (Fig. 9.2).

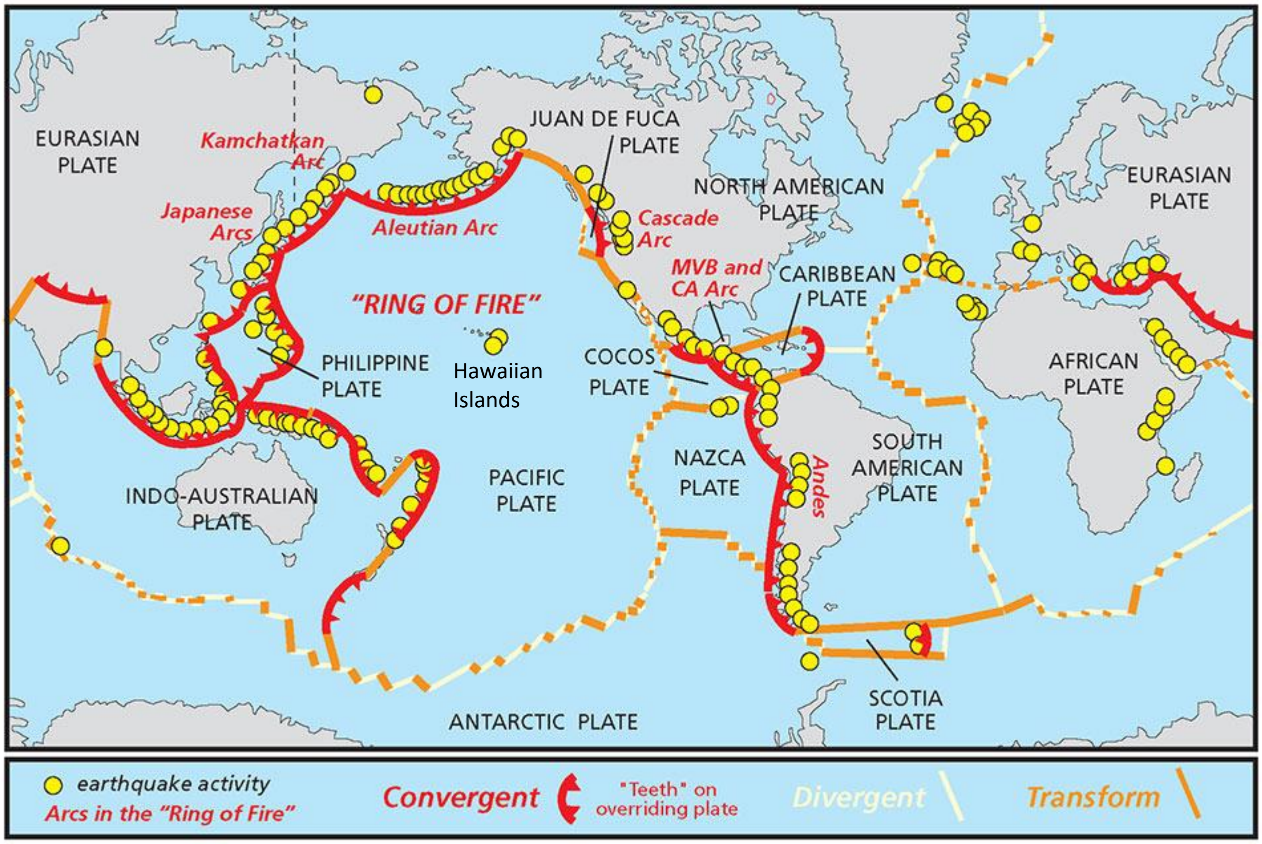

Figure 9.2 *– Earth's tectonic plates and types of plate boundaries (NOAA, 2021).*

9.4.2 Lithospheric Plate Boundaries

When lithospheric plates move relative to each other, there are three different types of boundaries that can occur (Fig. 9.2). When plates move away from each other, it creates a *divergent plate boundary*. When plates move towards each other, it creates a *convergent plate boundary*. Finally, there are also *transform boundaries,* where two plates slide by each other along a strike-slip fault zone. Plate boundaries can occur in the middle of the ocean, along the edges of continents, or in the middle of continents, depending on the type of boundary and how long it has been active.

Divergent Plate Boundaries

When soft asthenosphere rises by convection to the base of the lithosphere, it splits the lithosphere apart to form a *divergent plate boundary* (Fig. 9.3). Most divergent plate boundaries occur in the middle of the ocean as mid-ocean ridges, where two plates with oceanic crust at their surfaces are separating from each other. The *mid-ocean ridges* (MOR) are broad symmetrical mountain ranges that have a narrow central valley called a *rift valley* where the crust is splitting apart. As splitting occurs in the rift valley, it reduces pressure in the upper mantle, which triggers the partial melting of upper mantle rocks. The resulting magmas are basaltic and rise to the floor of the rift valley where they erupt on the sea floor to form *lava lakes* and *pillow basalts* that fill the rift valley. *Pillow basalts* are *subaqueous* eruptions of basalt in which fresh extruding lava develops in the shape of a pillow and forms a crust that splits and allows another pillow-like mass to extrude (Fig. 9.4). This type of eruption at mid-ocean ridges generates new crust and lithosphere with a basaltic composition. These eruptions far beneath the ocean surface are passive, and not the explosive types of eruptions that occur along convergent boundaries. If you were on a boat sailing across the ocean directly over one of these eruptions, you would not know it. The earthquakes produced by plate movement at the mid-ocean ridges tend to be relatively small because the movement at divergent boundaries creates tension (stretching) rather than compression. Tension creates normal faults and does not allow high stresses to build. Iceland is one of the few places on the planet where a mid-ocean ridge is spreading above sea level. This accounts for Iceland's frequent volcanic eruptions of basaltic lava, geysers, and many waterfalls dropping off the sides of the rift valley.

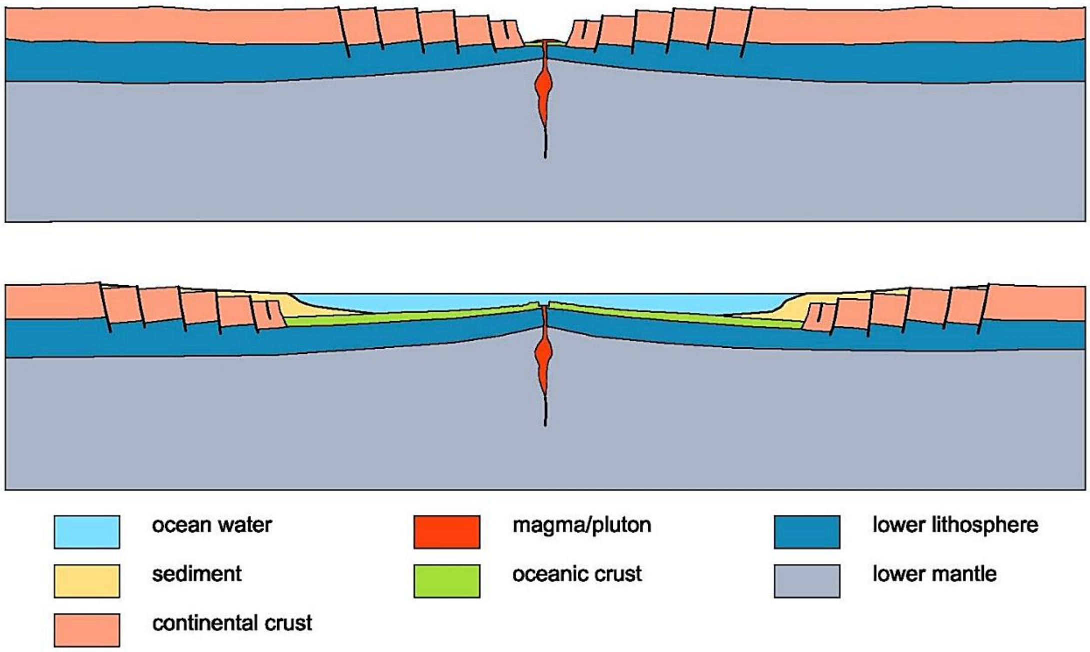

Figure 9.3 *– Cross sectional views of divergent plate boundaries. Note that the lower part of the lithosphere is also the upper part of the mantle. Splitting of a continent along a divergent plate boundary (above) eventually leads to the development of an ocean basin (below) as plates on both sides continue to move away from each other and new ocean crust is created. Sediment also accumulates along the edges of the continents.*

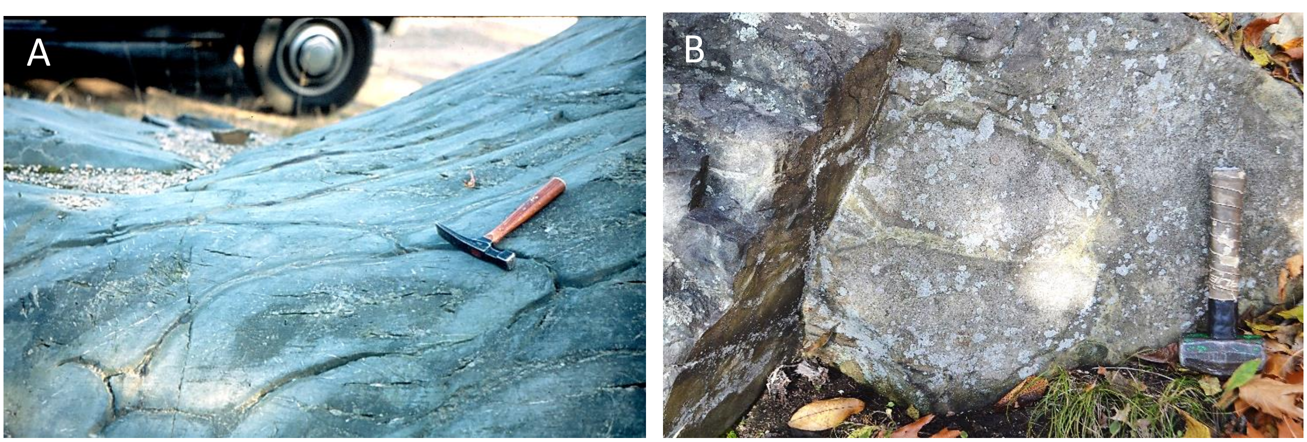

Figure 9.4 *– Pillow basalts in ancient rock formations. Rock hammers for scales. A) Archean pillow basalts at Bishop Corners, Ontario. Cracks have formed along pillow boundaries, which are a light-colored seam. Thick pillow rims weather to a lighter color, while pillow cores are darker and coarser. Image courtesy of James Hume. B) Late Proterozoic pillow basalt at Stoneham High School in Stoneham. These pillows are relatively small as compared to most pillows. The rims are dark and between the pillows is a light-colored, epidote-rich rock like in Ontario. The pillow cores weather to a light color and are coarser than the dark rims. This rock formation also occurs in the Fells, but no pillows have been found there.*

Divergent plate boundaries can also occur on continents as large *rift valleys*. As spreading occurs along theses boundaries, the rift valleys get deeper and wider and will sometimes form large linear lakes. These rift valleys accumulate sediment eroded from the surrounding high terrane adjacent to the valley and are interspersed with basaltic volcanic rocks. The best modern examples are the East African Rift Valleys (Fig. 9.5). Ancient examples of this situation are the sequence of Mesozoic red sedimentary rocks and basaltic lava flows that occur in the Connecticut River Valley in Massachusetts and Connecticut. As spreading continues, a rift valley will get longer and wider such that it eventually attaches to the ocean, as is the case with the Red Sea between Africa and Saudi Arabia (Fig. 9.6). It then becomes an infant ocean basin. Eventually, it will evolve into a large ocean basin (Fig. 9.3) as has occurred with the Atlantic Ocean.

Figure 9.5 *– Google Earth image of Lakes Tanganyika and Malawi (also known as Lake Nyasa) in East Africa. These lakes formed as a result of rifting of the crust.*

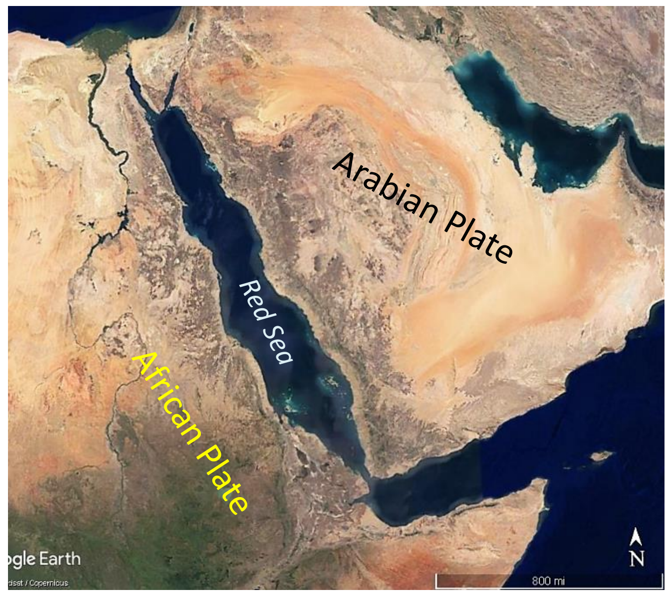

Figure 9.6 *– Google Earth image of the Red Sea where the Arabian Plate has now rifted away from the African Plate.*

Convergent Plate Boundaries

When soft asthenosphere sinks at the base of the lithosphere, it moves two plates towards each other and can drag one of the plates beneath the other at a *convergent plate boundary* or *subduction zone* (Figs. 9.7 and 9.8). This causes enormous compression and, as a result, the earthquakes generated at subduction zones tend to be much larger than at mid-ocean ridges. At subduction zones, only plates with oceanic crust at their surface can get subducted because continental crust is too buoyant to descend beneath another plate. Subducted basaltic crust will eventually reach a depth where it melts and forms magma. The magma may then rise and lead to the formation of plutons and volcanic eruptions. Rising basaltic magma, because of its high temperature, will melt rocks in overriding basaltic or continental crust, producing a wide range of magma compositions from sialic to basaltic. Since the magmas produced at subduction zones are partly derived from water-saturated sediment and rock at the surface of oceanic crust, they have abundant *volatiles* (water and gases) dissolved in them. Therefore, subduction zone eruptions tend to be explosive, and often emit large volumes of magma as airborne pyroclastic material. The

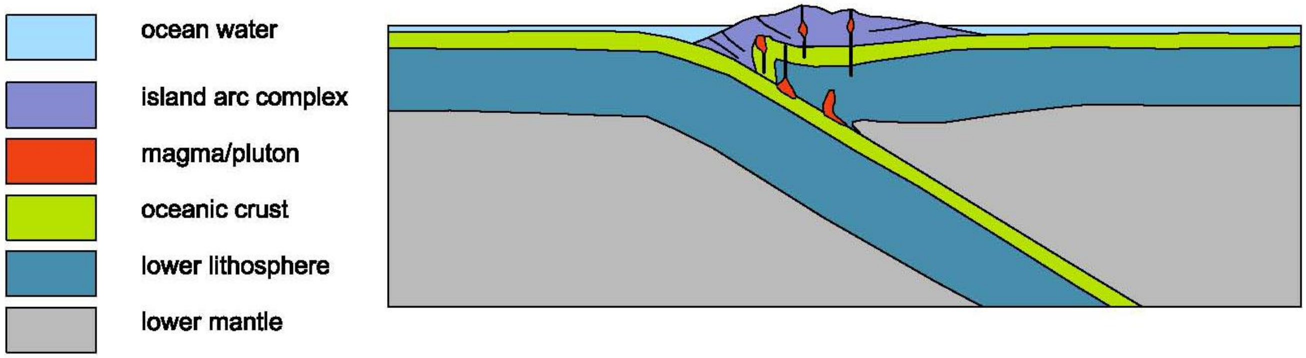

Figure 9.7 *– Cross-sectional view of an island-arc convergent plate boundary or island-arc subduction zone, also simply known as an island arc. Note that the lower lithosphere is also the upper part of the mantle. Descent of an oceanic plate leads to the production of magmas and volcanic activity. This is what is happening along the Aleutian Arc in Alaska as the Pacific Plate moves northward and descends beneath an oceanic plate in the Bering Sea (Fig. 9.2).*

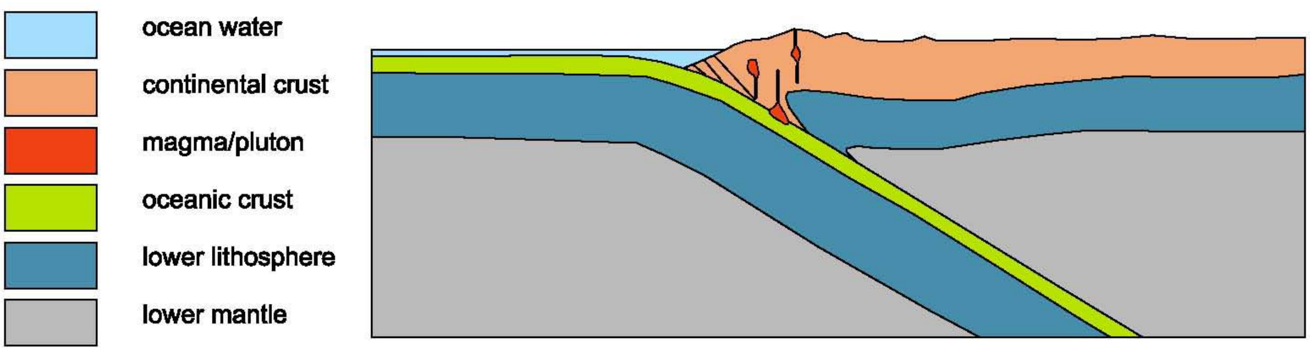

Figure 9.8 *– Cross-sectional view of a continental-margin convergent plate boundary or continental-margin subduction zone, also simply known as a continental arc. Note that the lower lithosphere is also the upper part of the mantle. Descent of an oceanic plate leads to the production of magmas and volcanic activity. This is what is happening along the west coast of South America where the Nazca and Antarctic plates are descending beneath South America (Fig. 9.2).*

large, memorable eruptions that have occurred in historic times, such as Tambora (1815), Krakatau (1883), Pinatubo (1991), and Mt. St. Helens in the U.S. (1980), were all at subduction zones. The perimeter of the Pacific Ocean is largely rimmed by subduction zones, and as a result is surrounded by large explosive volcanoes and a zone of powerful earthquakes. This is known as the “Ring of Fire” (Fig. 9.2).

Subduction zones can occur in the middle of the ocean where two plates with oceanic crust at their surfaces converge to form an *ocean-ocean subduction zone* (Fig. 9.7). Where one plate disappears beneath the other, it drags the ocean floor down and forms a long linear submarine valley called an *ocean trench*. Ocean trenches are the deepest features of ocean basins. In fact, the deepest place in the ocean, at about 10,900 m (35,761 ft) below sea level, is the Challenger Deep in the Marianas Trench, adjacent to the Marianas Island chain in the southwest Pacific Ocean. This is deeper than the tallest mountain, Mt. Everest, rises above sea level. When the down going plate eventually reaches a depth where it melts, magma rises to form volcanoes that line up parallel to the trench, like at the Marianas Island chain, forming an arc of volcanic islands. Convergent plate boundaries of this type are also called *island-arc subduction zones* or more simply, *island arcs*.

Subduction zones can also form where a plate with oceanic crust descends beneath a plate with continental crust, forming a *continental-margin subduction zone* (Fig. 9.8). In this case, there is an ocean trench just offshore from the continent, and volcanoes will erupt from beneath the leading edge of the continent, forming a *continental arc*. A prominent example occurs along the west coast of South America where the Nazca and Antarctic plates descend beneath South America and subduction is responsible for the Andes Mountains and Andean Trench (Fig. 9.2). In addition to volcanic activity, compression along the coast of South America generates many large earthquakes, and has uplifted, tilted, and deformed rock formations. Regional metamorphism is also altering rock formations at depth beneath the continent. With uplift and erosion, metamorphic rocks, as well as intrusive igneous rocks, formed from rising magmas, will eventually be exposed at the land surface. The sialic/felsic intrusions and volcanic rocks of the Fells were mostly formed in a continental arc environment.

Transform Plate Boundaries

The third type of plate boundary is a *transform boundary*, where one plate slides by another at a strike-slip fault zone (Figs. 9.2 and 9.9). Transform faults not only occur as long plate boundaries but also occur as shorter boundaries associated mid-ocean ridges. Mid-ocean ridges are not continuous features, but rather are composed of segments where rifting and spreading occur. Mid-ocean ridges occur as segments to accommodate the movement of plates on a curved (spherical) surface, where the plates are actually rotating on a global scale. The segments are separated by transform faults (see the Juan de Fuca Ridge on Fig. 9.9). The transform faulting occurs only between the two offset ridge crests because, outside this zone, there is no horizontal displacement between two parts of a single plate that are moving in the same direction. Few large earthquakes are generated along mid-ocean ridges, but when they do occur, they are usually at the transform boundaries.

In some places, transform faults can be very long, as with the San Andreas Fault in southern California (Fig. 9.9). The San Andreas Fault represents a long transform boundary with a large right-lateral displacement, in which the Pacific Ocean floor moves to the northwest relative to North America. This transform fault is not a single fault, but a fault zone with many smaller faults. The irregular pattern of this fault zone allows large stresses to build, which are then released in the large earthquakes for which the San Andreas Fault is infamous.

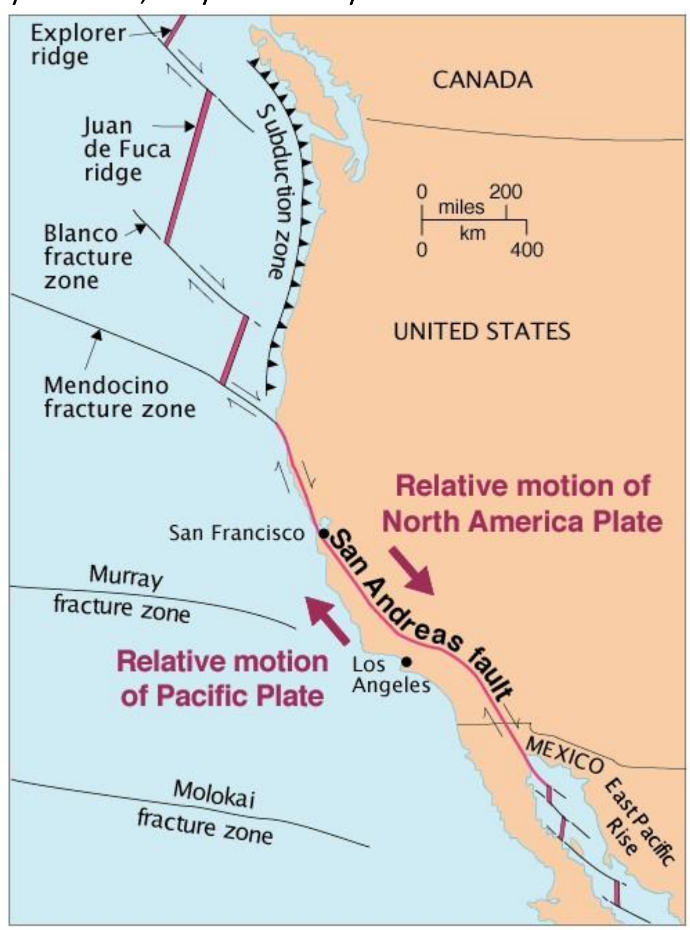

Figure 9.9 *–The San Andreas Fault in California is a transform fault boundary between the Pacific and North American plates (USGS, 2021). Note the smaller transform (strike-slip) faults that form offsets between different mid-ocean ridge segments like at the ends of the Juan de Fuca Ridge and between segments of the East Pacific Rise in the Gulf of California.*

9.4.3 Hot Spots and Mantle Plumes

What we have learned so far suggests that plate boundaries are the only places where volcanic activity can occur. However, plates sometimes rest over *hot spots* in the mantle that can generate magmas, which rise upward and reach the surface. In these places, the lower mantle is abnormally hot, and rising heat along with rising mantle material, called a *mantle plume*, will cause melting of the upper mantle and lower crust to produce plutons and volcanic eruptions. This can occur beneath plates that have either an oceanic or continental crust. The most widely cited example of an oceanic hot spot is the Hawaiian island chain (Fig. 9.2). The hot spot currently sits beneath the large island of Hawaii at the southeastern end of the island chain. It produces large volumes of basaltic magma that have piled onto the crust over the last few million years to form a large shield volcano. As the Pacific Plate has moved northwest across the hot spot, outpourings of basaltic lava have built new volcanic islands. As a volcano moves to the northwest beyond the hotspot it becomes extinct, and a new eruptive center forms over the hot spot. Over tens of millions of years this has led to the development of a chain of volcanic islands that are progressively older to the northwest. An example of a hot spot beneath a continent is the eruptive center of Yellowstone National Park. In this case, eruptions have produced a mixture of rocks ranging from mafic to sialic volcanic rocks, and some very explosive eruptions. In New England there are plutonic igneous rocks associated with a hot spot that was active in the Mesozoic Era. Jurassic age plutons in the White and Ossipee Mountains of New Hampshire (see Fig. 4.5 in Chapter 4) were formed by this ancient hot spot beneath continental crust.

9.4.4 What About The Rock Cycle and Tectonics?

How does plate tectonics change our planet? Basaltic rock of the ocean floor, which formed by volcanic activity at the mid-ocean ridges, slowly moves toward subduction zones. Along with the marine sediment that accumulates on the oceanic crust, the basalt gets destroyed during subduction, i.e., it gets metamorphosed and eventually returns to the mantle. Uplift of overriding plates at subduction zones causes the erosion of weathered rocks, which leads to the generation of sediment and sedimentary rocks. In the subsurface at subduction zones, metamorphism is changing rock formations and deforming subducted igneous and sedimentary rocks. Extreme conditions of pressure and temperature will cause some rocks to melt and form rising magmas, which will eventually form new igneous rocks. Rising magmas also cause contact metamorphism. We have seen how plate tectonics works on a large scale. Thinking back to the rock cycle (see Fig. 2.11, Chapter 2), you will remember that rocks can be uplifted, eroded, metamorphosed, and when reheated can generate new magmas. These processes are all the result of plate tectonics. As you can see, not only is plate tectonics alive and well, but so is the rock cycle as the plate tectonic system transforms rock formations into new ones, either following complete rock cycle loops or taking shortcuts. In other words, rocks get recycled by plate tectonics. In looking at the rocks in the Fells, it is easy to realize that the plutonic and volcanic rock units, and metamorphism of some units, have tremendous meaning in terms of ancient tectonic activity.

Just imagine what it would be like if Earth lost its interior heat (it's going to happen eventually!) and there was no active plate tectonic system. If plate tectonics shut down, there would be no more subduction, no uplift of mountains, and very few earthquakes and volcanoes. New oceanic crust would not be generated at the mid-ocean ridges. Existing mountains would eventually get torn down by erosion, and the continents would eventually be leveled to a low elevation by weathering and erosion. The gases emitted by volcanic activity, which significantly contribute to the greenhouse effect and warmth of the atmosphere, would be turned off. We might have to put more carbon dioxide in the atmosphere to prevent a catastrophic cooling of the planet to unlivably low temperatures, like on Mars. The situation described above of a tectonically inactive Earth is not likely for at least billions of years, so we have some time to prepare, and nothing to worry about right now. Our species (*homo sapiens*) may even be extinct or have evolved into different species by the time all this occurs.

9.4.5 Splitting and Assembling Continents

The global study of rock units has made it clear, that at times in the past, there were *supercontinents* that were assembled over long periods of time and later rifted apart by tectonic activity into smaller continents. Evidence for this today is some of the same evidence recognized by Alfred Wegener in the early 1900's. We can also now apply a fundamental understanding of how plate tectonics operates. The continents have outlines that fit together and rock formations and fossils on separate continents match each other. So how does this happen?

The splitting of supercontinents occurred as a result of rifting, perhaps triggered by strings of hotspots where plumes of mantle material rose by convection under a large continent. Rifting of a continent, if maintained, will first lead to a large rift valley, and eventually an ocean basin (Fig. 9.10). Relatively recent areas of rifting have occurred in and around Africa. In east central Africa are the

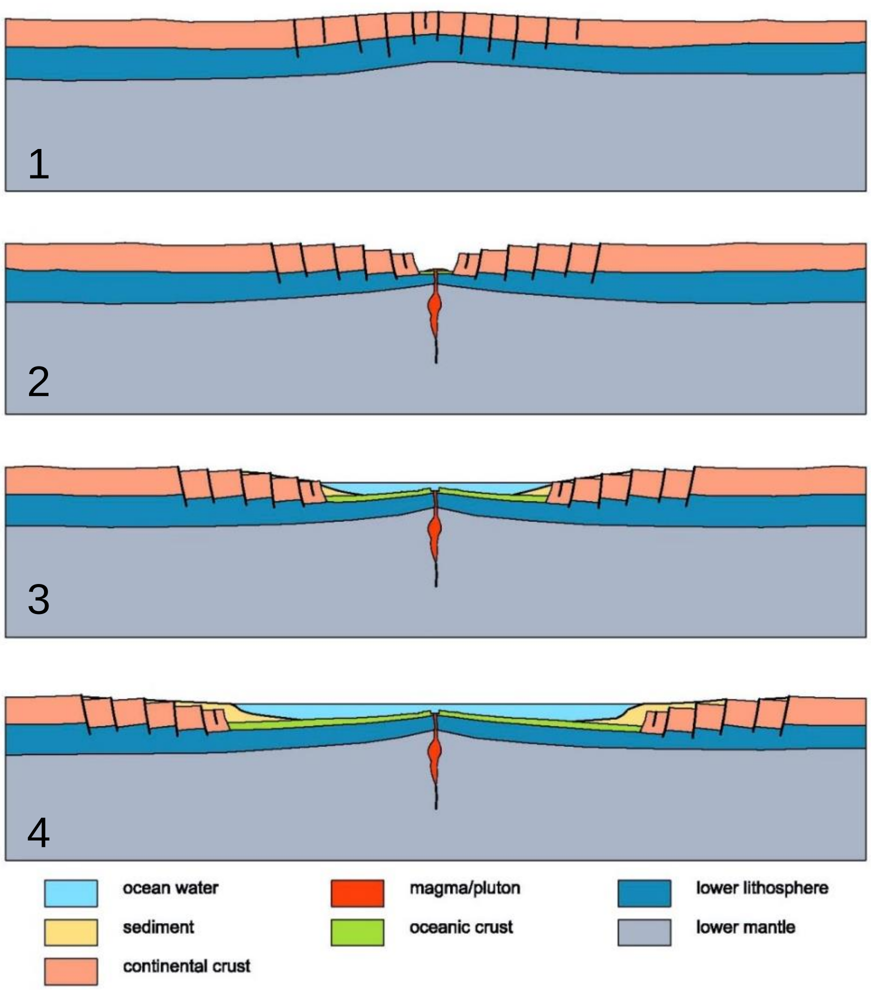

Figure 9.10 *– Stages in the development of a rift and ocean basin. Note that the lower lithosphere is also the upper part of the mantle. 1) Splitting of a continent begins with tensional fracture systems that eventually generate normal faults above areas where the asthenosphere (lower mantle) is rising due to upward mantle convection. 2) Initial splitting of a continent leads to the development of a rift valley bounded by normal faults, as occurs in East Africa today (Fig. 9.5). 3) Widening of a rift basin leads to the development of a narrow ocean basin, as occurs in the Red Sea today (Fig. 9.6). 4) Continued widening of a narrow ocean basin leads to a large ocean basin with thick packages of sediment accumulating along the edges of continents, as occurs today along the edges of the Atlantic Ocean.*

East African Rift Valleys that today are deep, wide valleys, which are still filling with sediment and volcanic rocks in an area where the crust has split apart. Some of these valleys are so deep they are occupied by lakes, such as Lakes Tanganyika and Malawi (Fig. 9.5). The Red Sea is another example of a rift where the Arabian Plate has pulled away from the African Plate (Fig. 9.6). This rift is still active and has spread to the point of becoming a narrow arm of the Indian Ocean. Over 10's of millions of years, it will continue to grow into a large ocean basin, like what has happened to the Atlantic Ocean since the Mesozoic Era. In the case of the Atlantic, the large supercontinent of Pangea, which existed back in the Paleozoic Era, rifted apart into separate continents that today are North America, Eurasia, Africa, South America, and Antarctica. India also rifted away from Pangea as a separate continent that later collided with Asia (see below).

We know that supercontinents can split apart, but how do they form, or assemble, in the first place? The assembling of supercontinents occurs when the ocean floor between two continents gets gobbled up by subduction, eventually leading to a continental collision (Fig. 9.11). This is precisely what has happened over the last 30 million years or so as India, which was a separate continent, collided with Asia. The resulting compression and uplift in this collision zone led to the development of the Himalayas. In this case, the subduction zone that dipped beneath Asia has become plugged with continental crust that is too light to be subducted, and India became a part of Asia.

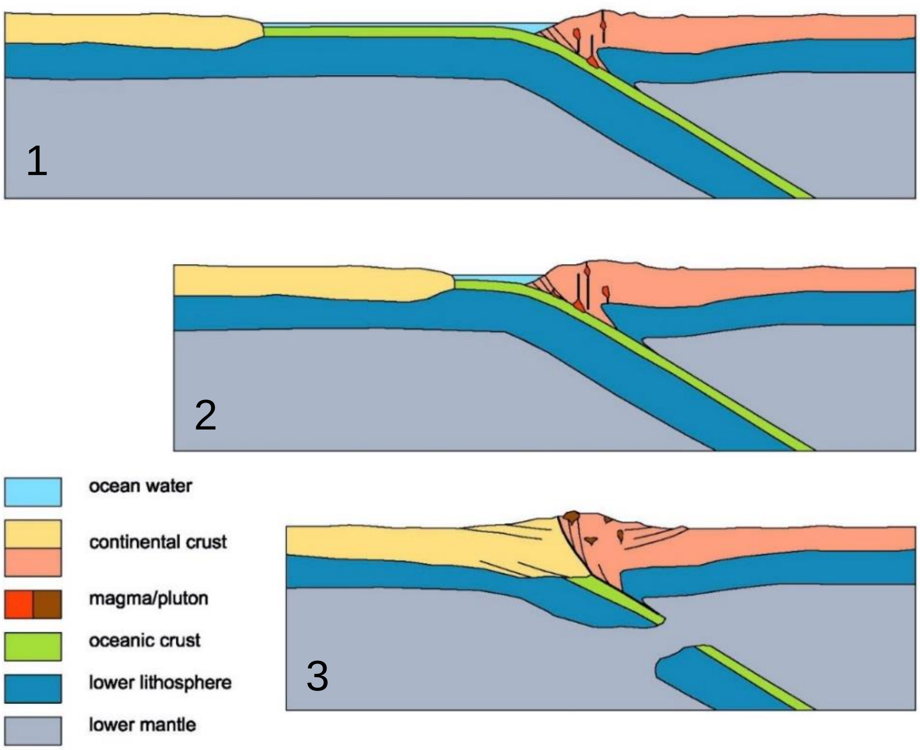

Figure 9.11 *– Idealized collision of two continents at a subduction zone and closure of an ocean basin. Note that the lower lithosphere is also the upper part of the mantle. 1) A wide ocean basin becomes narrower as oceanic crust is destroyed at a subduction zone. 2) A continent approaches a subduction zone as ocean floor gets subducted. 3) Collision of two continents as an ocean basin is completely subducted. At depth, the subducted plate detaches and sinks into the asthenosphere, where it is consumed in the lower mantle part of the asthenosphere.*

It is also possible for small continents, or *micro-continents*, and island arcs to get added, or *accreted* to a continent (*accretion*). Micro-continents can be small areas of continental crust torn from previous supercontinents by rifting, or they may be small masses of volcanic rock created at island arc subduction zones, or ocean hot spots. Micro-continents and island arcs have densities that are too low to allow them to get subducted and so instead they get accreted onto the edge of a larger continent (Fig. 9.12). Subduction may be halted and resumed on the opposite side of the accreted land mass. In the case of an island arc, it may also eventually lead to a change in the direction of subduction (Fig. 9.13). We call this activity *terrane accretion* or *terrane docking*. Southern Alaska is a good example of a continental area that has had many *terranes* dock on to it, with each terrane bounded by faults and having a unique set of rocks. The history of the Appalachians in New England is also one of collisions in which continents, micro-continents, and island arcs were accreted to North America. The Fells sits in one of these terranes called the Avalon terrane, or Avalonia, which was accreted in the Silurian Period. Pieces of Avalon can be traced north of New England through the Maritime Provinces of Canada to Newfoundland. Avalon originally became a micro-continent when it rifted off what is today West Africa. Accretion is a common process along the edges of many continents, where the margins of the continent are composed of multiple deformed terranes. Rocks in a continent's interior generally escape the effects of collisional tectonics and are not deformed. This is the reason that sedimentary rocks in the midwestern U.S. generally have only gentle dips.

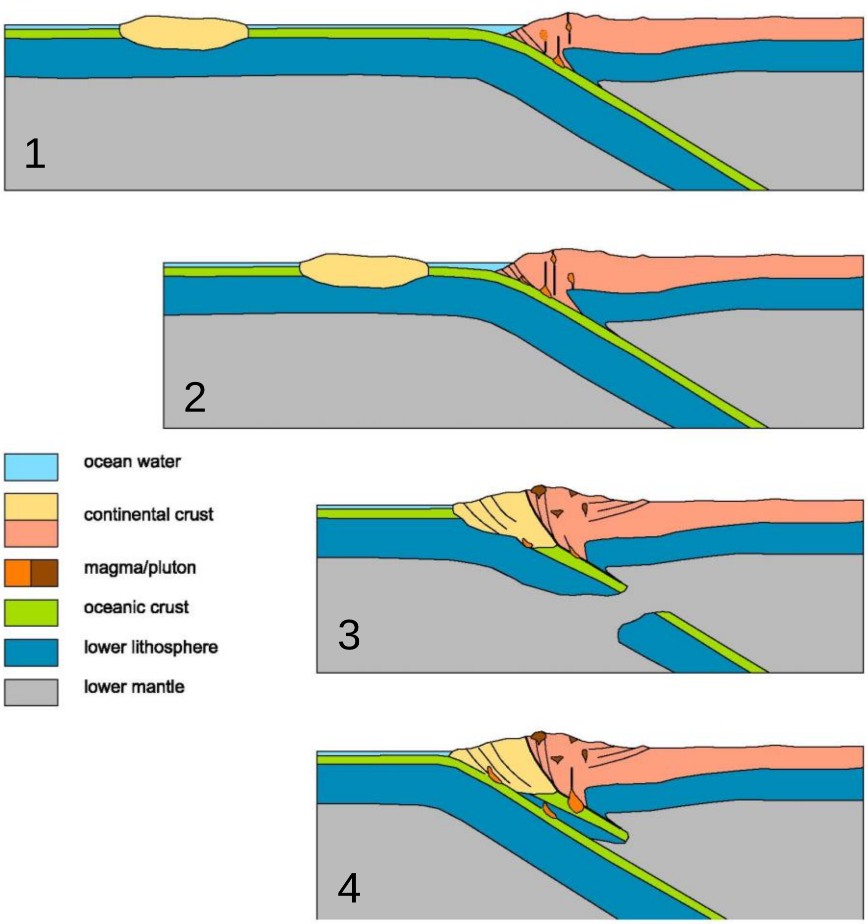

Figure 9.12 *– Idealized micro-continent accretion along a continental margin subduction zone. 1) Micro-continent in the middle of a large ocean basin. 2) Micro-continent approaches the coast of a continent. 3) Collision of the micro-continent with the continent at a subduction zone. The down going plate detaches. 4) A new subduction zone may form on the opposite side of the micro-continent, which is now accreted to the continent and becomes a terrane.*

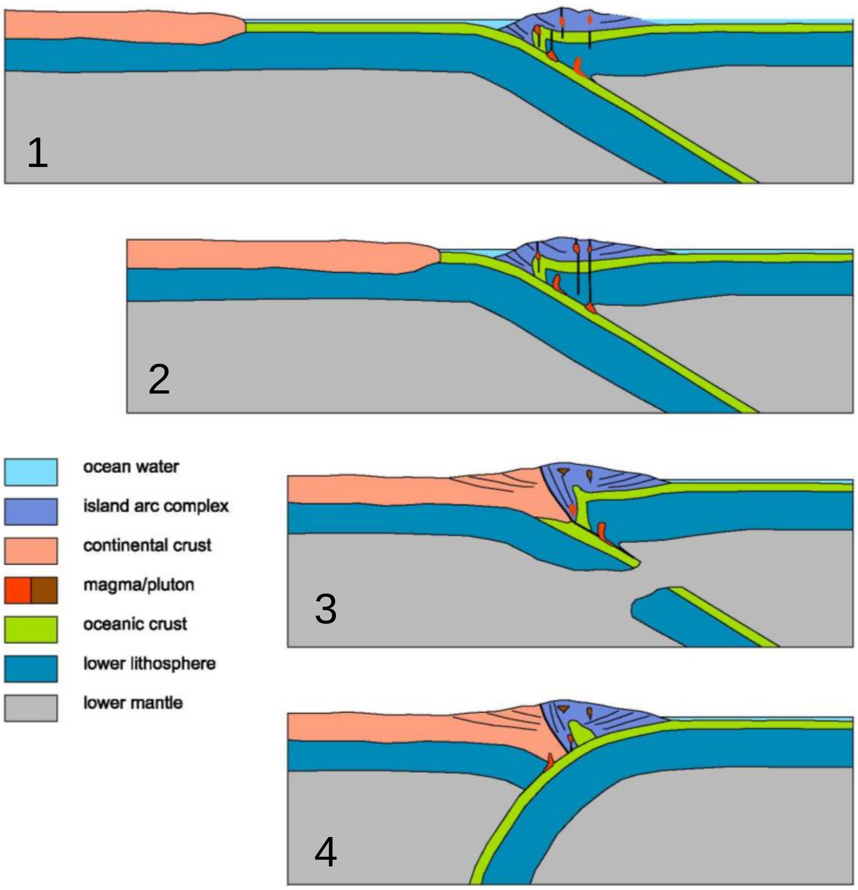

Figure 9.13 *– Idealized collision of an island arc subduction zone with a continent. 1) A wide ocean basin separates a continent and island arc. 2) A continent has arrived near the island arc. 3) Collision of the continent with the island arc. The down-going plate becomes detached. 4) A new subduction zone may form on the opposite side of the island arc.*

The few examples of collisional tectonics given here are relatively simple examples. It can be much more complex than this with collisions involving two subduction zones or when plates approach each other at an oblique angle. Later deformation and metamorphism of previously accreted terranes also complicates the interpretation of the rocks we see exposed in a mountain belt today. The challenge for geologists is to properly decipher rocks to identify terranes and terrane boundaries and to understand their plate tectonic significance, eventually leading to a reconstruction of a continent's tectonic history.

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