



10. Continents: structure and history

We saw in chapter 3 that the Earth's outer shell – or lithosphere – is broken into plates moving relative to one another (plate tectonic). The lithosphere comprises the crust and the uppermost part of the upper mantle. The crust is subdivided into an oceanic and a continental crust. The former is heavier, thinner, and lies beneath the ocean floor which represents 2/3 of the Earth's surface. The latter is lighter, thicker, and makes up the continents. The oceanic lithosphere is produced at mid-ocean ridges and destroyed along subduction zones. Therefore, the oldest oceanic lithosphere present beneath the ocean floor is only 200,000,000 years old, which represents a relatively small fraction of Earth's 4.6-billion years history. On the other hand, continental rocks can be much older because they are too light to be subducted. Hence, only continental rocks can be used to study the evolution of our planets in the distant past (billions of years ago).

The edge of a continent facing a subduction zone is called an **active margin**. A good example is the west coast of South America. The edge of a continent that does not coincide with a convergent plate boundary is called a **passive margin**. The Atlantic Ocean is bordered by passive continental margins.

10.1. Structure of continents

The surface of continents can be subdivided into several broad regions which were formed at different times and by different tectonic processes. These broad regions are referred to as **tectonic provinces**:

Shields lie at the heart of continents and are their oldest components. Shields are composed of Precambrian crystalline rocks (metamorphic and igneous rocks). These rocks may have been deformed during the Precambrian but remained undeformed during the Phanerozoic. Examples of shields are the Canadian Shield, the African Shield, and the Australian Shield.

Platforms consist of sedimentary layers overlying the Precambrian basement and displaying relatively minor deformations (e.g. North American Platform, Siberian Platform). Shields and platforms consist of the most stable parts of continents called **cratons**. Cratons have not been significantly affected by mountain building episodes (orogenies) during the Phanerozoic. However, they may have experienced slow vertical movements due to mantle convection beneath the lithosphere or to the growth and retreat of ice sheets (see epeirogeny, section 10.3). Large areas can be uplifted due to heating by mantle upwelling beneath the lithosphere*. More locally, magmatic intrusions migrating upward can warp the overlying rock layers. Circular regions of the lithosphere which have been uplifted are called **structural domes** (e.g. Cincinnati Arch, USA; Richat Dome, Mauritania). Rock layers in a structural dome dip away from its center. On the other hand, regions of the lithosphere which cools and contracts are affected by subsidence and are

* Note that vigorous upwelling may cause crustal extension which in turn may lead to rifting and the formation of a sedimentary basin. An ocean may even form if ocean-floor spreading is initiated and maintained (see the Wilson cycle, section 10.2.3).



called **thermal subsidence basins*** where abundant sediments can accumulate (e.g. Michigan Basin, USA; Amadeus Basin, Australia). Unlike domes, rock layers in a thermal subsidence basin dip toward its center. Intracratonic basins are economically important because they contain large reservoirs of coal, oil, and natural gas.

Phanerozoic orogens are mountain chains situated along the margins of cratons and produced by continental collision or ocean-continent subduction. The Appalachian mountains (eastern USA) and the Ardennes (central Europe) are examples of past orogens related to mountain building episodes that are no longer active. On the other hand, the Alpine-Himalayan orogen and the Andes are examples of ongoing mountain building episodes (see slides for more details).

Regions of **extended crust** are regions where the last episode of tectonic deformation is an extension. Passive margins are regions of extended crust because they are formed after rifting and the subsequent opening of an ocean basin (e.g. the passive continental margins bordering the Atlantic Ocean).

10.2. Growth of continents

Two main processes contribute to the growth of continents: one that involves vertical transport of matter (**magmatic addition**) and one that involves horizontal transport of matter (**accretion**).

10.2.1. Magmatic addition

Abundant magma is produced at subduction zones through water-induced melting of the subducting oceanic plate and the mantle directly above it (see chapter 6). After being produced, the magma rises toward the surface and either crystallizes in a magmatic chamber or is spewed out during volcanic eruptions. In Japan, the occurrence of many granitic intrusions and active volcanoes attests of the importance of magmatic addition for the formation of the Japanese Archipelago.

10.2.2. Accretion

A continent can also grow larger when a portion of the neighboring plate merges with the continent. Accreted blocks of tens to hundreds of kilometers in size are called **terrains** (or terranes). The geology and age of these accreted terrains may be very different from those of the adjacent continental rocks. Japan is formed of a juxtaposition of accreted terrains parallel to the subduction zone and bounded by major faults. The age of terrains decreases toward the trench. They are composed of portions of the oceanic crust as well as volcanic and sedimentary rocks carried by the subducting plate. An example of accreted material observed in Japan is the siliceous sedimentary rock called chert (see chapter 7). It is a layered rock composed of tiny shells made of silica which have accumulated on the deep ocean floor. Another example of accreted material in Japan is reef limestone. A prominent example is the Akiyoshi Terrain in Yamaguchi Prefecture which is

* Note the analogy with the thermal subsidence experienced by passive margins and permitting the thick accumulation of sediments on continental shelves.



composed of atoll reefs ranging in age from Early Carboniferous to Middle Permian.

Continental growth by accretion does not only happen in subduction zone. Probably the most obvious example of continental growth by accretion is when two continental landmasses merge during continental collision (e.g. collision between India and Eurasia forming the Himalayan mountain chain). Less obvious but nonetheless of fundamental importance is the accretion of blocks moving along a transform fault. This process has shaped the west coast of North America. For example, the terrain named Wrangelia now forming part of Alaska and western Canada originated from the southern hemisphere and has travelled 5000 km north to finally merge with North America.

10.2.3. **The Wilson cycle**

The margins of cratons have experienced several episodes of mountain building throughout Earth's history. Meanwhile ocean basins have opened and closed repeatedly. Since the surface area of the Earth is constant, plate divergence must be compensated by convergence. Divergence is linked to the opening of ocean basins whereas convergence leads to the closure of ocean basins and continental collision. An idealized picture of the evolution of continents and ocean basins is given by the **Wilson cycle**. Once during the Phanerozoic all the continents merged into a single supercontinent called **Pangaea** about 250 million years ago. Pangaea was not the first supercontinent and will not be the last. The Wilson cycle begins with the break up of a supercontinent through rifting and the opening of an ocean basin followed by a subduction leading to continental collision and ultimately to the assembly of a new supercontinent. Rifting of the new supercontinent marks the beginning of the next cycle. It takes several 100s of millions of years to break-up and then reassemble a supercontinent.

10.3. Vertical movements of the continental lithosphere

Continental collision results in a great deal of crustal deformation (faults and folds), metamorphism, and crustal thickening leading to the formation of magma at the root of mountain chains. But the continental lithosphere can also be affected by vertical movements with little or no deformation. These vertical movements of the continental lithosphere are referred to as **epeirogeny**. There are two main causes of epeirogenic movements: (1) glacial rebound and (2) the heating/cooling of the lithosphere.

10.3.1. **Glacial rebound**

The weight of large glaciers can bend the continental lithosphere. When these glaciers melt, the lithosphere rises progressively. High-latitude regions of the northern hemisphere like Scandinavia are now being uplifted following the melting of the thick ice sheet that covered these regions during the last ice age (110,000-12,000 years ago).



10.3.2. *Heating/cooling of the lithosphere*

Upwelling of mantle rock beneath the continental lithosphere causes uplift. When heated, the lithosphere expands and its density decreases which makes it “float” higher on the asthenosphere. Moreover, the pressure of the rising mantle rock may also push the lithosphere upward. Uplift of the lithosphere may lead to the formation of structural domes, already mentioned in section 10.1. The Basin and Range Province in western USA is a result of the heating of the lithosphere by mantle upwelling inducing uplift and crustal extension accompanied by faulting.

Conversely, cooling of the continental lithosphere causes subsidence. This process leads to the formation of thermal subsidence basins where abundant sediments can accumulate (e.g. Michigan Basin, passive margins of continents).

Thermal subsidence and uplift are processes occurring over millions of years. Conversely, glacial rebound is faster and occurs over thousands of years.

10.4. Origin of the resistance of cratons

Cratons are composed of very old crustal rocks. Evidence for continental crust 4 billion years old has been found in Canada. Cratons are very stable. Cratons may contain rocks which have remained undeformed since the Archean era (3.9-2.5 billion years ago). On the other hand, the margins of these stable cratons have been affected by multiple orogenic episodes. How can we explain the extraordinary resistance and stability of the cratons?

The continental lithosphere is light and cannot be subducted. On the contrary, rocks of the oceanic crust are constantly being recycled in the mantle. Hence, the rocks beneath the ocean floor are less than 200 million years old whereas those of the continental crust can be billions of years old. But how have cratonic rocks resisted tectonic deformation for so long? The answer probably lies in the nature of the lithosphere beneath the cratons. Cratons are characterized by a particularly thick lithosphere with a rigid lithospheric mantle that acts like a bumper and prevent cratons from being crushed during collisions. The thick and rigid lithospheric mantle beneath cratons is referred to as the **cratonic keel**. There is much speculation concerning its structure and composition. The low geothermal gradient beneath cratons suggests that cratonic keels are colder than the surrounding mantle rocks. If their composition was the same as the mantle material around, they would tend to sink because of their lower temperature, hence their greater density. One hypothesis suggests that cratonic keels are composed of a material that is less dense than the mantle around. Evidence supporting this hypothesis comes from a rock called kimberlite. This rock is formed during a violent volcanic eruption which brings material lying deep in the lithosphere up to the surface. These kimberlites are found within cratons and contain fragments of the cratonic keel (together with diamonds). The composition of these fragments is that of a peridotite. The rock is similar to “standard” mantle peridotite except it contains less iron and less garnet and consequently has a relatively lower density.