7. Sedimentary rocks

Sediments and sedimentary rocks cover 90% of Earth's surface. They are essential to our understanding of the origin and evolution of life because they are the rocks containing fossils. They are the rocks from which coal, oil, and natural gas can be extracted. Limestones are sedimentary rocks made of $CaCO_3$, a major component of cement and many other industrial products. Moreover, salt deposits (evaporites) are a primary source of NaCl (table salt) and other chemical compounds. Sedimentary rocks are also often associated with reservoirs of groundwater.

7.1. What are sediments?

Examples of ***sediments*** include beach sand, gravels of riverbeds, and fine particles of desert dunes. These are sediments that have not yet been transformed into sedimentary rocks. They are still soft, ***unlithified***. We will see in section 7.3 how soft sediments become a solid ***lithified*** rock.
Sediments can be classified into two broad categories: solid matter and ions in solution.
Examples of solid sedimentary particles: (1) fragments of minerals and rocks, (2) hard parts of organisms (e.g. mollusk shell, coral skeleton), and (3) organic matter (e.g. plant leaves, wood).
Examples of ions in solution: Na^+, K^+, Cl^-, Ca^{2+}, CO_3^{2-}…

7.2. How do sediments form?

7.2.1. Weathering

Sediments form by destruction of preexisting rocks. The destruction process is called ***rock weathering***. Two types of weathering can be distinguished: (1) ***physical weathering*** and (2) ***chemical weathering***.

- *Physical weathering*

Major physical weathering agents are ***wind*** (eolian weathering), ***water***, and ***ice*** (glacial weathering). The dust carried by the wind abrades the surface of rocks. The polished surface of stones in arid regions such as the Moroccan desert or Mars illustrates well the process of eolian weathering. Water is a powerful weathering agent. The destructive action of waves along shorelines is a good example. Receding shorelines due to wave action are common around the world. Given enough time, rivers can cut through hard rocks and form ***V-shaped valleys*** with steep flanks (e.g. Grand Canyon, USA). Ice is also a powerful agent of rock destruction. Glaciers are like rivers of ice which can carve large ***U-shaped valleys***. During the last ice age, the northern polar ice cap was much more extensive than today. Regions of North America and northern Europe that are now ice-free still bear the marks of glacial weathering. Ice can also destroy rocks through ***frost wedging***. This weathering process results from the fact that the volume of water increases when it freezes. If liquid water fills the fractures in a rock and then freezes, the ice

expands and widens the fractures, which can in turn break the rock.
Plants may also participate in rock physical weathering through ***root wedging***. Plant roots in the fractures of a rock force them to open further as the plant grows, which may lead to rock dislocation (***biophysical weathering***).

- *Chemical weathering*

Rainwater and CO_2 combine in soils to form the weak carbonic acid H_2CO_3. In contact with carbonate rocks, the carbonic acid dissolves $CaCO_3$. The products of this reaction, Ca^{2+}, HCO_3^- and CO_3^{2-}, are transported by rivers to the ocean. In the ocean, the ions Ca^{2+} and CO_3^{2-} can be used by marine organisms to build their calcareous hard parts (see related slide). Dissolution of $CaCO_3$ is responsible for the formation of caves and other dissolution features (***karst***).

The process of $CaCO_3$ dissolution does not necessarily require the action of living organisms. However, the active dissolution of $CaCO_3$ by organisms happens as well. For example, some bivalves are capable of dissolving $CaCO_3$ and bore holes in which they live (***biochemical weathering***).

Like carbonate rocks, silicate minerals can react with acidic groundwater. The products of this weathering reaction include dissolved ions (e.g. Ca^{2+}, Si^{4+}, Fe^{2+}, HCO_3^-, CO_3^{2-}, and H^+) and clay minerals. Ions are transported by rivers to the ocean and can be used by marine organisms to build hard parts made of $CaCO_3$ and SiO_2.

The weathering of silicate rocks results in a net removal of CO_2 from the atmosphere over millions of years. This slow removal of CO_2 is balanced by an equally slow input of CO_2 from volcanoes and hot springs. This is an important mechanism for regulating Earth's climate at geological timescales －over millions of years－ (see related slides for further explanation).

7.2.2. Erosion and transport

The removal of sediments from their source area is called ***erosion***. Erosion is carried out by the same flowing agents causing rock physical weathering: ***wind***, ***water***, and ***ice***. Sediments are then transported by these same agents to the site of their deposition. If there is no active removal, sediments may simply move away from their source area by gravity. The downslope movement of sediments by gravity is called ***mass wasting***.

7.2.3. Sediment deposition

Solid sedimentary particles carried either by ***wind*** (eolian transport), ***water*** or ***ice*** (glacial transport) will at some point be deposited. Deposition of a sedimentary particle transported by wind or water occurs when the wind/water speed decreases and is no longer capable of carrying the

particle in question.
In water, strong currents with a speed of >50 cm/s can carry all sizes of particles up to boulder-size sediments (>25.6 cm). Moderately strong currents with a speed of 20-50 cm/s can carry sand (62.5 µm-2 mm) and smaller particles. Weak currents (<20 cm/s) can carry silts and clays (<62.5 µm). Note that we consider here only sediments consisting of fragments of rocks and minerals which generally have roughly the same density. We don't consider material with a very low density like wood.
Sediments carried by ice are deposited when the ice melts. The nature and morphology of these deposits is typical of glacial deposits and can be recognized in the field. Glacial deposits are useful indicators of past climate as they are indicative of great accumulations of ice.
In the case of mass wasting, deposition is controlled by the nature of sediments (e.g. size, shape) and the slope angle.

For ions in solution, deposition occurs either abiotically or biotically. An example of abiotic deposition is the precipitation of minerals in a saline lake by evaporation to form a type of rock called ***evaporite*** (see also chapter 4). Biotic deposition of ions in solution is the process of biomineralization. Corals, for example, use Ca^{2+} and CO_3^{2-} present in the ocean to produce their skeleton made of $CaCO_3$.

7.3. How do sedimentary rocks form?

How are soft sediments, say a beach sand for example, transformed into hard sedimentary rocks, in this case sandstone?
As soft sediments accumulate, they are progressively buried under younger layers of sediments. The set of physical and chemical changes which happen to soft sediments after their burial is called ***diagenesis***. The result of diagenesis includes the transformation of soft sediments into hard rock. This transformation is called ***lithification***. Lithification results from two main diagenetic processes:

(1) ***Compaction***: as sediments are buried deeper and deeper under younger sedimentary layers, the pressure and temperature increase. Sediments are squeezed, compacted. The space between sediment grains gets smaller. Dissolution also occurs where sediment grains are in contact as a consequence of increased pressure.
(2) ***Cementation***: dissolved ions precipitate in the open space left between sediment grains. Common mineral cements include calcite ($CaCO_3$), hematite (Fe_2O_3), and quartz (SiO_2).

Compaction and cementation contribute to reduce the rock ***porosity***. Porosity is a measure of the amount of open space (or ***pores***) in a rock.

7.4. Properties of sediments and sedimentary rocks

The characteristics of sediment grains can provide information about the environment in which they have formed. Fundamental properties of sediments and sedimentary rocks include:

Grain size: Since mineral and rock fragments have generally roughly similar densities, grain size gives a good indication of the strength of the current that has transported them. For example, a rock composed of sand-sized sedimentary grains is indicative of a stronger current relative to a rock composed of clay particles.

Sorting: Sorting is a measure of the variation in grain size within a sediment or sedimentary rock. If all sediment grains have approximately the same size, the sediment is well sorted. If the sediment is made of a mixture of small grains and large grains, the sediment is poorly sorted. The level of sorting provides information about the type of current which transported the sediment. If the strength of the water current is constant over time, the size of sediment grains deposited tends to be uniform. If the strength of the water current is variable, the size of grains deposited varies accordingly.

Grain morphology: During transport sediment grains are abraded and become rounded. Therefore, the roundness of sediment grains provides information about the distance over which sediments were transported. Greater distances of transport result in greater roundness.

7.5. Sedimentary basins and depositional environments

Large amounts of sediments are deposited in depressions of the Earth's crust. Large depressions in which abundant sediments accumulate are called ***sedimentary basins***. The largest basins are the ocean basins. Their size and morphology is controlled by plate tectonics. An example of a large continental sedimentary basin is the East African Rift which corresponds to a divergent boundary where two plates are pulled apart. The rift contains many lakes in which sediments accumulate. In the future, if the plates keep moving in opposite directions, the continental lithosphere will stretch and get thinner; the rift will grow larger and deeper, leaving more space for sediments to accumulate. The downward movement of the lithosphere as it is stretched and thinned is called ***subsidence***. Subsidence is accentuated by the weight of sediments pushing down the lithospheric plate. Seawater will ultimately flood the rift and seafloor spreading will initiate, transforming the continental rift into an ocean basin. The two continental margins that are moving apart progressively cool as they are moving away from the source of heat, i.e., the mid-ocean ridge. As it cools, the lithosphere becomes denser and progressively sinks, creating space for sediments to accumulate. This is a particular type of subsidence called ***thermal subsidence*** (see corresponding slide). It enables the accumulation of thick piles of sediments along continental margins and is a key process in the formation of ***continental shelves***.

There are many different kinds of ***depositional environments*** (e.g. deserts, lakes, rivers, deltas, beaches, organic reefs, continental shelves, deep sea…) which belong to either one of the following three categories: (1) *continental*, (2) *shoreline*, and (3) *marine environments*. Each depositional environment is characterized by a unique set of physical, chemical, and biological processes. The nature of sediments deposited in these environments is determined by these processes. For example, the deep sea is not affected by strong, sustained currents like shallow-water environments. The deep sea is thus generally characterized by the deposition of fine-grained sediments (mostly mud). Deep-sea organisms are also different from those living in shallow waters. Let's take another example: coral reefs. Coral reefs are associated with tropical and subtropical climates. They grow in shallow water and their rigid framework consists mainly of coral skeletons and crustose coralline algae.

Useful indicators of past depositional environments are ***sedimentary structures***. When one looks at an outcrop of sedimentary rocks, the most conspicuous sedimentary structure is often the ***bedding*** (see related slides for illustrations). Sedimentary beds can be a few cm thick to several m thick. Bedding results from changes in sedimentation affecting grain size and/or sediment composition. The equivalent of bedding at the mm scale is called ***lamination***. Bedding is useful to study tectonic deformation (folds and faults) because sediment layers are usually deposited originally flat and horizontal. Other remarkable sedimentary structures are ***ripple marks*** (see related slides for illustrations). Ripple marks are a common feature of sandy beaches. The geometry of ripple marks preserved in sedimentary rocks provides information about the transport agent (wave vs. wind, unidirectional vs. bidirectional, current direction). ***Burrows*** are sedimentary structures too. They represent one example of sedimentary structures resulting from the activity of living organisms. Other biological sedimentary structures include traces left by marine organisms moving on the seafloor.

In conclusion, geologists are able to reconstruct past environments based on the characteristics of sedimentary rocks (characteristics of mineral grains and cements, fossil content, and sedimentary structures). By studying successions of sedimentary rocks of different ages exposed on Earth's surface or uncovered by drilling, it becomes possible to study environmental changes over geological time scales. The fossil content of the rocks can also be used to gain an understanding of the nature and mechanisms of biological evolution.

7.6. Types of sedimentary rocks

7.6.1. Siliciclastic sedimentary rocks

Siliciclastic sedimentary rocks are composed of fragments of rocks (lithic fragments) and minerals. The composition of these rocks is dominated by silicate minerals, like quartz or feldspar.
Siliciclastic rocks are classified according to either grain size (coarse = ***conglomerate***, medium = ***sandstone***, fine = ***shale***) or grain composition (e.g. ***arkose*** = sandstone containing at least 25% of a mineral called feldspar).

7.6.2. Biochemical sedimentary rocks

Biochemical sedimentary rocks are usually formed by the accumulation of the hard parts of marine organisms (e.g. bivalve shells, coral skeletons…). A beach deposit consisting of fragments of bivalve shells (***bioclasts***) is an example of biochemical sediments. Coral reef framework is another example. Bivalve shells and coral skeletons are made of $CaCO_3$ but another major biomineral is silica (SiO_2). Marine microorganisms, such as diatoms and radiolarians, producing tiny shells made of silica can accumulate on the deep seafloor and form siliceous rocks called ***diatomite*** and ***radiolarite***.
The constituents of biochemical sedimentary rocks are generally formed by the direct biological precipitation of $CaCO_3$. But biological activity can also induce precipitation of $CaCO_3$ indirectly. Photosynthetic microbial mats called stromatolites are a good example. The equilibrium between dissolved calcium and carbonate ions in the sea and the mineral calcium carbonate can be written as follows: $Ca^{2+} + 2HCO^{3-} \leftrightarrow CaCO_3 + CO_2 + H_2O$. The reaction of photosynthesis is $6H_2O + 6CO_2 + \text{sunlight} \rightarrow C_6H_{12}O_6 + 6O_2$. Photosynthesis consumes CO_2 whereas calcification releases CO_2 in the environment. Consequently, the photosynthetic activity of stromatolites tends to promote calcification by shifting the equilibrium state of the carbonate system to the right ($Ca^{2+} + 2HCO^{3-} \rightarrow CaCO_3 + CO_2 + H_2O$). Stromatolites can form rigid pinnacles and these structures owe their rigidity to the early cementation induced by photosynthesis (see related slide for illustration).

7.6.3. Chemical sedimentary rocks

Chemical sedimentary rocks are formed by mineral precipitation which does not involve biological activity. For example, minerals precipitating at the bottom of a saline lake subject to intense evaporation. The accumulation of minerals in this kind of setting produces a rock called ***evaporite***. One common constituent of evaporites is NaCl (halite or table salt).

7.6.4. Organic sedimentary rocks

Organic sedimentary rocks are formed by accumulation of organic matter (e.g. plants). Usually, organic matter is quickly decomposed by bacterial activity. However, in particular settings which are characterized by a high rate of accumulation of organic matter and low levels of oxygen, organic matter can be preserved and form thick accumulations. An example of such accumulation is ***peat***. Peat forms in wetlands by accumulation of plants. Low oxygen concentrations prevent the complete decomposition of organic matter. As peat is buried under younger sediments, the increase in pressure transforms peat into ***lignite*** (compressed peat). At greater depth, pressure and temperature increases further and lignite is converted into ***coal***. Coal is a complex mixture of mineral matter and organic carbon compounds. Increasing further the pressure and temperature transforms coal into ***anthracite***. Anthracite has a higher carbon content than coal. Coal and anthracite have a different chemical composition than peat and lignite. Heat and pressure have modified their composition and texture. These rocks have undergone some degree of metamorphism (see next chapter).