Petrography of sedimentary rocks

Introduction

Rocks can be classified into three main groups: igneous or magmatic rocks, sedimentary rocks, and metamorphic rocks. Igneous rocks result from the cooling and crystallization of magmas, originating from either the mantle or the melting of metamorphic rocks. Metamorphic rocks result from the alteration, by the action of heat and pressure, of igneous or sedimentary rocks, which come from the lithification by diagenesis of sediments. As these sediments come from the breakdown of sedimentary, metamorphic, or magmatic rocks, all these phenomena form a cycle called the geological cycle (Figure 01 and 02).

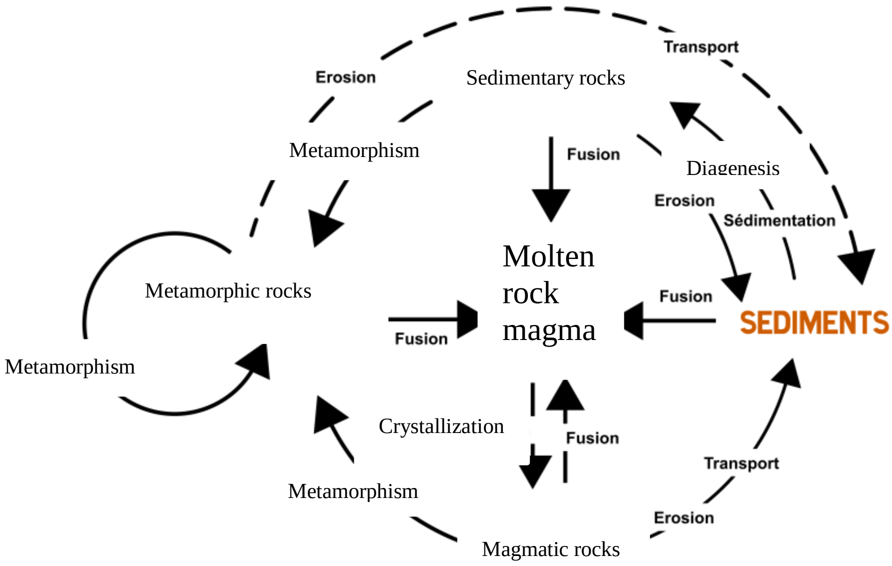


Fig. 01: The rock cycle (Bourque, 2005)

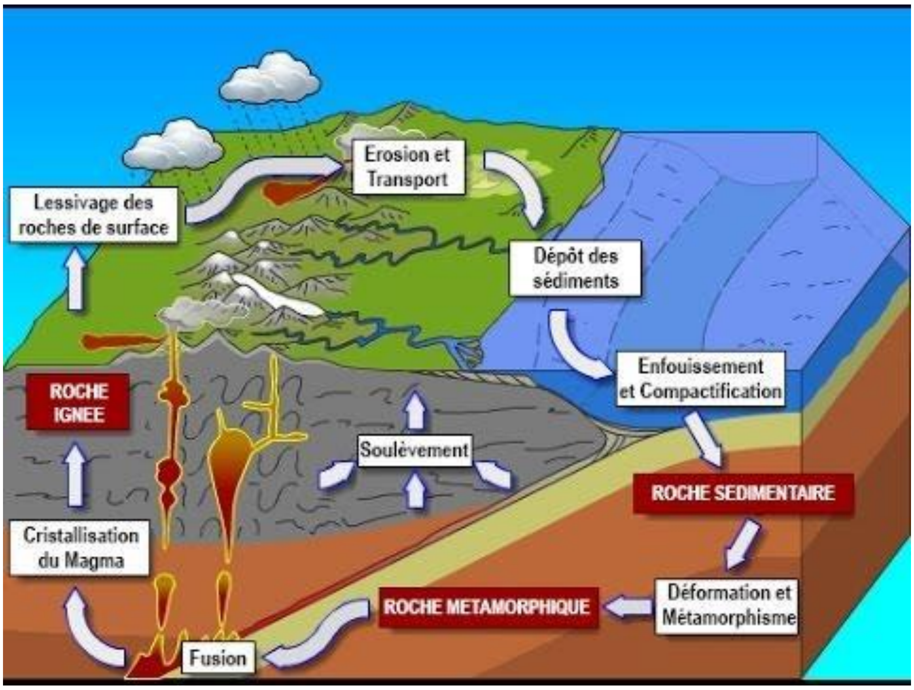


Figure 02: Schematic of sedimentary processes in the geological cycle

In the geological cycle, sedimentary processes include weathering, erosion, transportation, deposition, and diagenesis. Detrital sediments, formed from grains resulting from the breakdown of pre-existing rocks, transported and deposited in a sedimentary basin, represent the most evident illustration of this part of the geological cycle.

II- Genesis of sedimentary rocks

Four processes contribute to the formation of sedimentary rocks: the superficial weathering of materials generating particles, the transportation of these particles by rivers, wind, or ice to the deposition environment, sedimentation causing these particles to settle in a specific milieu to form sediment, and ultimately, diagenesis converting the sediment into sedimentary rock (Figures 03 and 04).

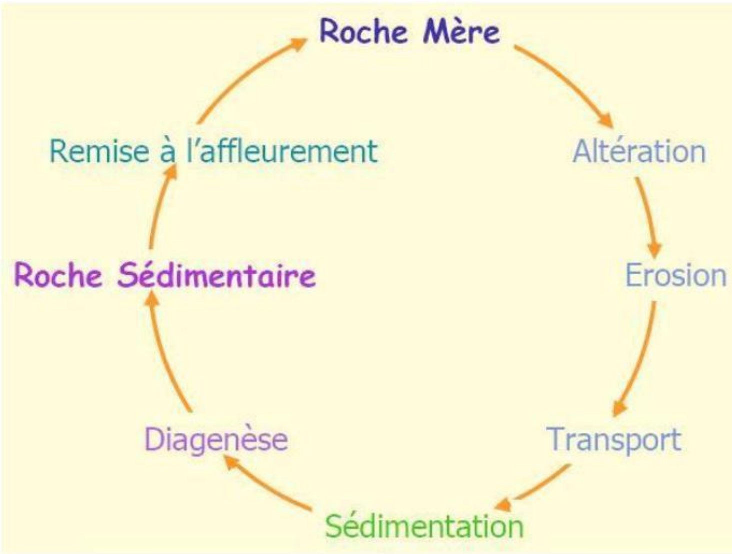


Fig. 03: The cycle of sedimentary rocks (Based on Emmanuel et al. 2007)

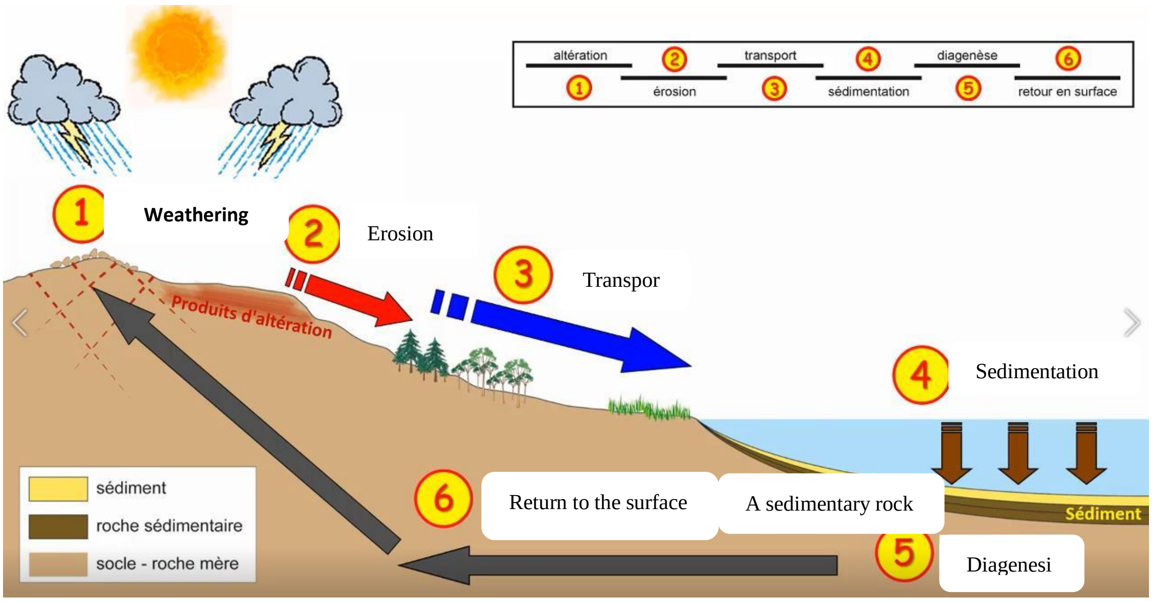


Fig. 04: Main processes of sedimentary rock formation

A) Weathering:

The processes of superficial weathering are three types: mechanical, chemical, and biological (Figure 05).

Fig. 05: Examples of the mechanisms of weathering

<u>Physical (or mechanical) weathering</u>

It is facilitated by the nature of the rock, the presence of fractures, and climatic variations. It is active in features like faults, joints, and pores. This weathering is caused by:

Cryoclasty (frost weathering): Corresponds to the cycles of freezing and thawing, which leads to the fragmentation of rock due to the pressures caused by the freezing and thawing of the water it contains. When the water freezes, it expands the fractures, causing the rock to break apart.

Thermoclasty corresponds to repeated temperature fluctuations, causing the fragmentation of rock. This process involves the expansion (during warm periods) and contraction (during cold periods) of minerals. Thermoclasty is prevalent in tropical deserts where temperature variations can exceed 50°C.

Decompression will gradually develop parallel joints on the surface of the soil.

Haloclasty: It corresponds to the fragmentation of a rock mass by salt crystals formed as a result of the evaporation of constituent water. These growing crystals exert strong pressure, resulting in the fragmentation of the rock.

Mechanical abrasion by detrital grains transported by the wind, water, or ice.

<u>Biological weathering</u>

involves plants and animals playing a significant role. This can be through the chemical action of compounds produced by organisms like plants or microbes, or the mechanical action of plant roots penetrating rock crevices, leading to its fragmentation. Animals also contribute by digging burrows, especially affecting less consolidated rocks.

Chemical weathering:

Water serves as the primary agent in this process. When in contact with CO_2, it behaves as a weak acid, dissolving chemical elements from the parent rock. The majority of reactions involved in this weathering process necessitate the presence of water and air.

Examples:

- Dissolution: $NaCl + H_2O = Na^+ + Cl^- + H_2O$
- Acid hydrolysis transforms silicate minerals into clays. Most of these minerals are only partially dissolved, resulting in a solution of silica and cations (K^+, Mg^+), along with a clay mineral.
- Transformation of primary minerals such as micas and feldspar into other mineral species (phyllosilicates like clays).
- Hydration: impacts rocks composed of elements capable of binding a water molecule within their structure; for instance, anhydrite converts to gypsum through hydration.
- Oxidation: primarily affects iron-rich minerals. When a mineral containing iron comes into contact with oxygen, it oxidizes to form another mineral. The presence of water accelerates this oxidation.

B)Erosion:

Erosion involves the mobilization of elements resulting from weathering, caused by wind, water, glaciers, and living organisms (bioerosion). This process depends on:

- Wind speed.
- Topography; steep slopes erode more quickly than gentle slopes.
- Presence of vegetation (acts as a protector).
- Grain size; blocks are less eroded than sand.

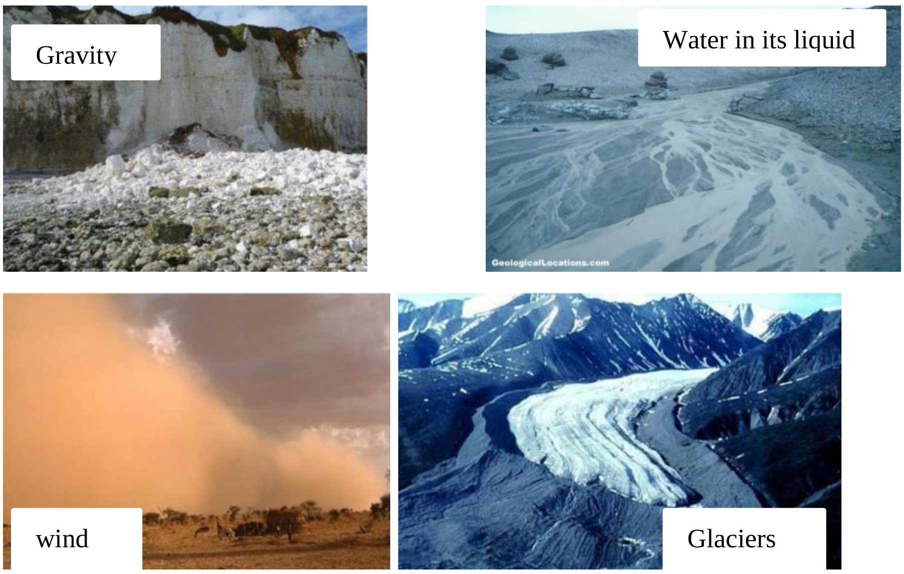


Fig 6 Examples of erosion and transport agents include

We distinguish:

B.1. Wind Erosion: This erosion, controlled by the wind, is significant in deserts, plains, and coastlines, and includes two processes: deflation and abrasion.

B.2. Fluvial Erosion: Water bodies like streams, rivers, and rivers can erode their substrates, transport, and accumulate sediments.

B.3. Glacial Erosion: The movement of glaciers leads to the detachment of blocks and the wear of these blocks due to the friction of particles of all sizes contained in the ice.

B.4. Marine Erosion: Waves and water currents are the main erosive agents in coastal environments.

C- Transport:

Apart from wind and ice, water is primarily responsible for transporting particles. Depending on the mode and energy of transport, the resulting sediment will exhibit various sedimentary structures: laminar stratification, oblique or cross-bedding, sorting by grain size, diverse markings at the top of layers, etc. Sedimentary rocks will inherit these structures. Particle transport can be very lengthy. Ultimately, all particles must end up in the ocean basin.

D) Deposition (or sedimentation):

Deposition occurs when the transport speed decreases until it becomes zero. All transported materials accumulate in a sedimentation basin, ultimately in the marine basin, to form a deposit. Sediments are deposited in successive layers whose composition, particle size, color, etc., vary over time depending on the nature of the sediments brought in. This is why sedimentary deposits are stratified and sedimentary rocks resulting from these deposits make up stratified landscapes like those in the Grand Canyon of Colorado, for example.

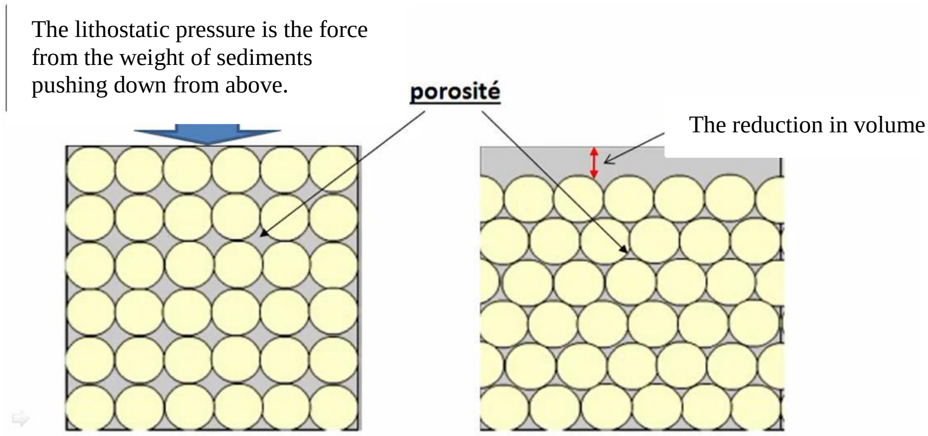
E) Diagenesis:

The formation of a sedimentary rock occurs through the transformation of sediment into rock under the influence of diagenetic processes. Diagenesis encompasses the physico-biochemical modifications that a sediment undergoes after its deposition in a sedimentary basin under conditions of "low" pressure and temperature prevailing in the subsurface. Diagenesis stops where metamorphism begins.

Diagenesis transforms loose sediment into consolidated rock through lithification. It is a relatively simple process: if the water flowing through a sediment, such as sand, is supersaturated with certain minerals, it precipitates these minerals into the pores of the sand, which then bind the sand particles together; this results in a sedimentary rock known as sandstone. The degree of cementation can be low, resulting in a friable rock, or it can be extensive, leading to a very solid rock.

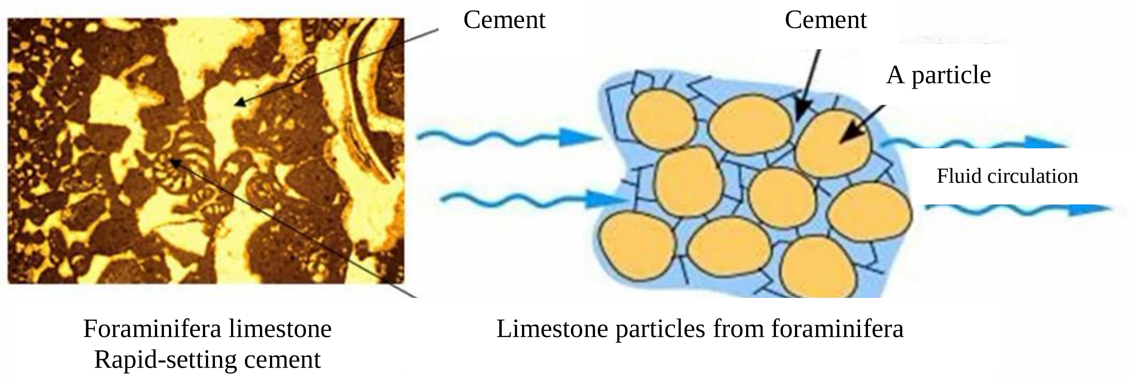
The diagenesis begins on the seafloor in the case of marine sediment and continues throughout its burial, as other sediments cover the deposit and gradually bring it deeper under several tens, hundreds, or even thousands of meters of material.

The diagenetic processes are varied and complex: they range from sediment compaction to cementation, involving phases of dissolution, recrystallization, or replacement of certain minerals. These processes include:
Compaction: It's when the pressure from above sediments increases, forcing water out of the sediments. This loss of water reduces pore size, causes a decrease in volume, and an increase in density. The grains move closer together and rearrange to withstand this pressure.



Process of Compaction

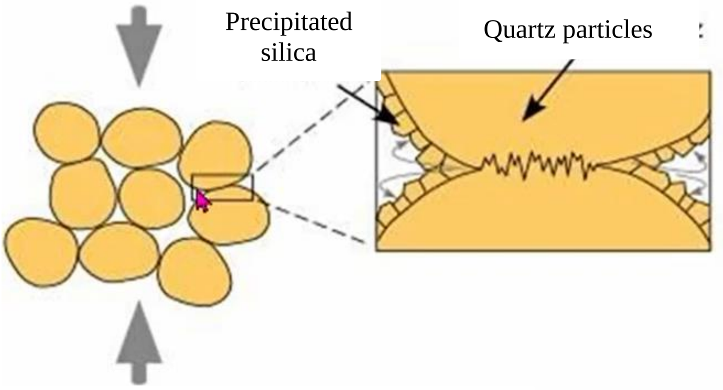
Dissolution: This process involves the dissolution of the chemical constituents of grains, sediments, or rocks into a solution, resulting in an increase in sediment porosity due to elevated pressure and the presence of aggressive waters in the pores (waters with high dissolved CO_2, high pH).
Cementation: It's when sediments or minerals precipitate among themselves, reducing porosity and increasing the size of the formed crystals. Water carrying ions or clay flowing through the deposit allows minerals to form between particles, sticking them together. Common cements include clay, quartz (or silica), and calcite. Cementation can occur early or late in the sediment's diagenetic history, either before the stacking of several meters of sediments (pre-compaction) or later when pressure on the particles is high due to sediment stacking.



Example of pre-compaction cementation (early cementation) (Bourque, 1997)

In the context of pre-compaction cementation (top diagram), fluids circulating within the sediment precipitate chemicals that effectively weld the particles together. The compaction of a sediment can result in cementation.

Consequently, the elevated pressure applied at the contact points between quartz particles in a sand triggers local dissolution of quartz, fluid oversaturation with silica, and subsequent silica precipitation on particle surfaces, thereby cementing them into cohesive structures.



Compaction and cementation

Recrystallization: Recrystallization is a process that results in a change usually seen as an increase in size or shape of a crystal without any change in chemical composition.

Example: The recrystallization of aragonite into calcite

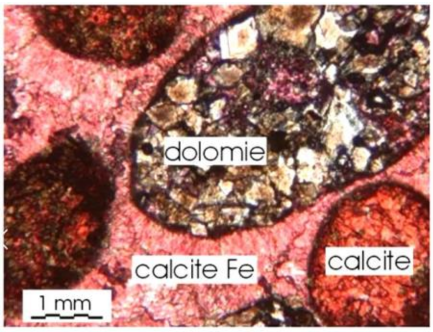
Aragonite (CaCO3) → Calcite (CaCO3)

(orthorhombic system) (rhombohedral system)

Replacement: Replacement is a diagenetic reaction where a mineral dissolves and another mineral precipitates in its place. This process usually occurs without any change in volume between the replaced mineral and the replacing one.

Example: Dolomitization (dolomitization of limestone)

Silicification (silicification of a fossilized wood)



Dolomitization of limestone
Substitution of calcium atoms with magnesium atoms

Fossilized silicon wood
Substitution of calcite atoms with silica atoms

Diagenetic evolution can schematically be divided into four successive phases of varying and increasing durations, during which various phenomena emerge, develop, and subsequently diminish (Figure 07).

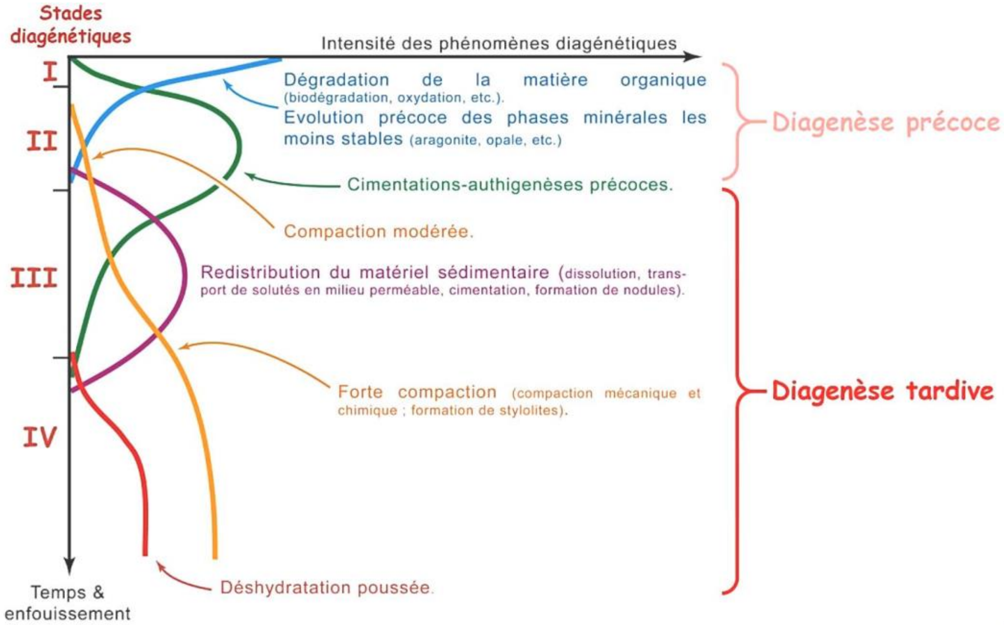


Fig. 07: Principal phenomena developed during diagenesis. The time scale follows a geometric progression from Phase I to Phase IV, representing varying burial depths depending on the nature of the different sedimentary constituents (in H. Chamley, 2000).

Biochemical diagenesis,

favored by microbial activity and linked to the early evolution of carbonates, silica, and potentially evaporites, characterizes Phase I. The majority of organic matter, particularly of animal origin, is decomposed, while aragonitic tests, the most fragile calcite tests, and a substantial portion of biogenic silica are dissolved. Concurrently, various gases are released (CO2, H2S, NH3). As an illustration, this initial phase affects common hemipelagic muds up to a maximum thickness of a few hundred meters. It is noteworthy that halmyrolysis refers to the alteration and pre-diagenetic crystallization processes occurring at the water-sediment interface.

In Phase II, the neoformation of sulfides, metal oxides, and various silicates (such as opal and zeolites), as well as the alteration of clay minerals, primarily occur. This phase takes place while interstitial water laden with dissolved products still circulates relatively freely within the sediment undergoing compaction. Clayey-limestone hemipelagites experience this phase over several thousand meters in thickness.

The process of cementation, leading to the filling of rock voids through ion migrations originating from distant sources, involves the precipitation of opal, quartz, calcite, dolomite, iron oxides, clay minerals, and more. It notably characterizes Phase III. This phase can occur at a depth of a few hundred meters for permeable rocks (sandstones) and much deeper for nearly impermeable clay rocks (around 10 km).

The concentration of chemical elements mobilized during the preceding phases around a nucleus, forming concretions, also characterizes Phase III. Nodules and concretions of marcasite, chert, sulfates, various carbonates, and phosphates are common examples of this process.

Intense dehydration begins in Phase IV and is accompanied by various recrystallizations, particularly near grain contact points where preferential dissolution occurs under pressure (resulting in the formation of secondary porosity or sutures like stylolites, cone-in-cone structures, etc.).

The rock, with its evolution dominated by pressure-dissolution phenomena, becomes more homogeneous as the residual voids are filled with newly formed minerals, whose ionic constituents have a predominantly local origin: salts, silica, quartz, feldspars, and clays.

During Phase IV, compaction intensifies significantly. Starting from the early stages of burial, it leads to a certain reorientation of particles (e.g., micas and clays) and various loading, expulsion, and shrinkage features within the loose sediment. This process results in a substantial reduction in porosity, flattening of fossils, imprinting of directly contacting grains, and the development of intricate fractures in deeper layers.

Metasomatism or epigenesis involves the replacement, under high pressure and often elevated temperatures, of certain minerals with more stable ones without a change in shape: late dolomitization and silicification, late iron mineralization, loss of volatile matter from fossil fuels, etc. After Phase IV, there is a gradual transition, influenced by increasing pressures and temperatures and through anchimetamorphic processes, to various metamorphic zones and ultimately anatexis. These latter phenomena gradually take over from diagenesis at significant depths (several tens of kilometers for very fine-grained rocks).

III- MAIN SEDIMENTATION ENVIRONMENTS

Reminder about the depositional environment:

The elements intended to form sediment are transported in a solid state or in solution. They then deposit or precipitate in a sedimentation environment.

A sedimentation environment is a basin (depression) where a set of chemical, physical, and biological factors are sufficiently constant to form a characteristic deposit; examples include fluvial, lacustrine, deltaic, and marine environments.

Thus, a geologist can reconstruct the conditions that prevailed in an ancient environment by examining the characteristics of its deposits. Understanding and mapping ancient sedimentation environments are fundamental aspects of paleogeography.

In the current natural world, we distinguish between continental environments, marine environments, and intermediate environments located at the boundaries of land and sea, known as mixed environments because they exhibit characteristics influenced by both land and sea proximity.

<u>A- Continental Environments:</u>

These types of environments are characterized by the absence of the sea. We can distinguish the glacial environment, characterized by the presence of water in its solid state; aquatic environments, where sedimentation is influenced by the presence of water and its level of agitation, subdividing into alluvial (fluvial) environments, lacustrine environments, and palustrine environments (saltwater lakes and marshes); terrestrial environments where sedimentation occurs dry or at least without constant water involvement (such as desert environments).

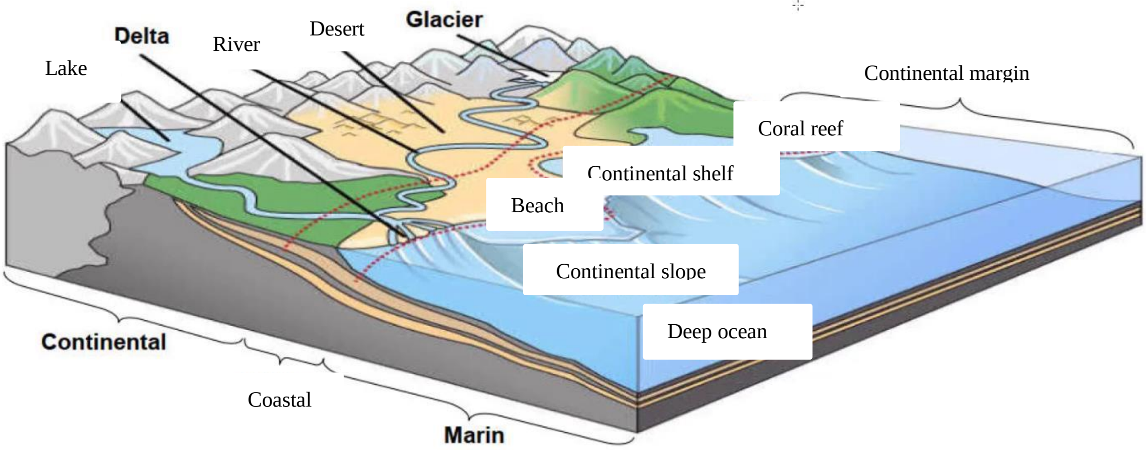


Fig.08: Main Sedimentation Environments

A-1. The glacial environment

The glacial environment is characterized by low temperatures that limit plant growth and animal life. Biological and chemical factors have negligible importance here, as the influence of ice and meltwater movements predominates, with meltwater often accumulating in very calm lakes. Glacial deposits are therefore entirely composed of clastic materials, ranging in size from huge boulders to the finest clays, with the boulders often showing characteristic scratches carved during transport by the ice. Deposits directly left behind by melting ice lack stratification and sorting. These are the various forms of moraines.

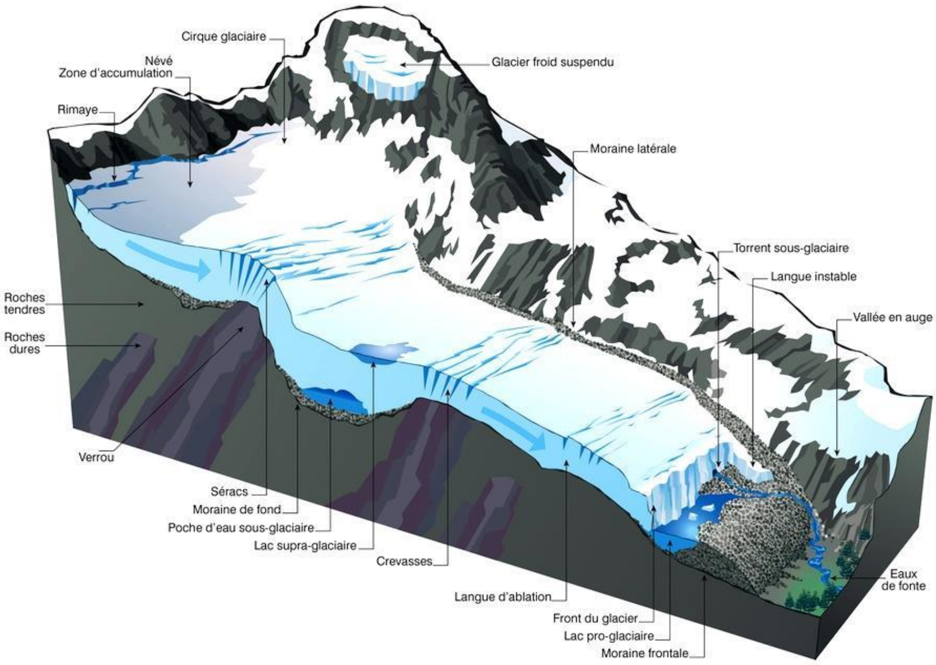


Fig.10: Block diagram of a valley glacier

Fig. 09: Frontal moraine of a glacier

A-2. The desert environment

In desert environments, sedimentation is primarily driven by the wind. Desert deposits consist mainly of pebbles, gravel, and sands of varying fineness, bearing the marks of aeolian erosion. The decrease in temperature promotes rock fragmentation through the freezing and thawing of water within them.

These deposits are often lens-shaped and may exhibit rapid, usually interlayered alternations of fine and coarse detrital material, desiccation cracks, ripples, aeolian dunes, and loess. (Loess: a loose, detrital sedimentary deposit, with particle sizes similar to silt (2 to 50 µm), mainly composed of silica (detrital quartz) and calcium carbonate ($CaCO3$). Of aeolian origin, deposited during cold and dry climatic phases, it covers vast areas and contributes to the formation of deep and fertile soils).

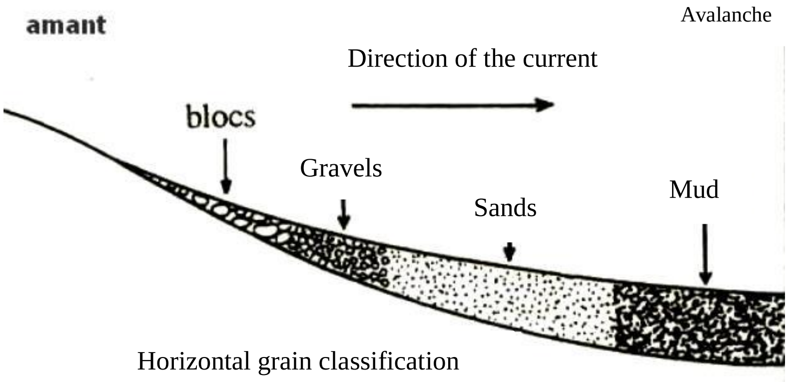
Fig. 11: Cross-bedded aeolian sandstone, Navajo (Colorado, United States)

Fig. 12: Mudcracks (desiccation cracks), Tassili N'Ajjer, Algerian Sahara

Aeolian sedimentary deposit Transverse dunes on the right and ripples on the left.

A-3. In a fluvial environment

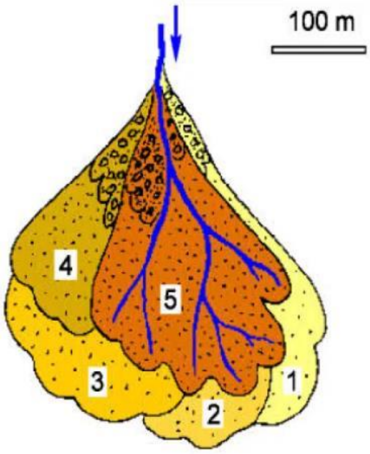
At high altitudes, near the source of rivers, the swift current allows for the erosion and deposition of large, angular elements. Downstream, erosion continues, causing the riverbed to deepen. Sedimentation occurs along the banks. Farther downstream, the decrease in current velocity leads to the deposition of finer elements, resulting in grain sorting or horizontal sedimentation of transported materials. The larger particles settle closer to the source where the current is strong, while the finer particles settle downstream where the current is weaker.

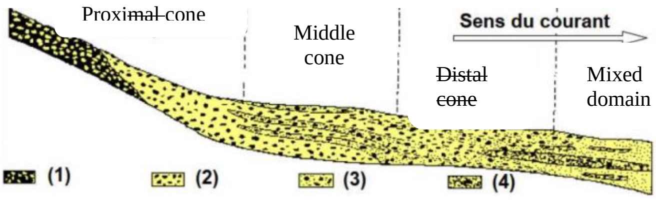


The materials of all sizes transported by the torrent can be temporarily deposited in the riverbed, but they are picked up during each flood to eventually be deposited when the velocity decreases, typically as the watercourse reaches a plain. This forms a lobed fan shape, known as an alluvial fan or a torrential cone of deposition.

Fig. 16: Simplified longitudinal section of an alluvial fan:

(1) Mudflows, (2) Pebbles, (3) Sands and gravels, (4) Silts



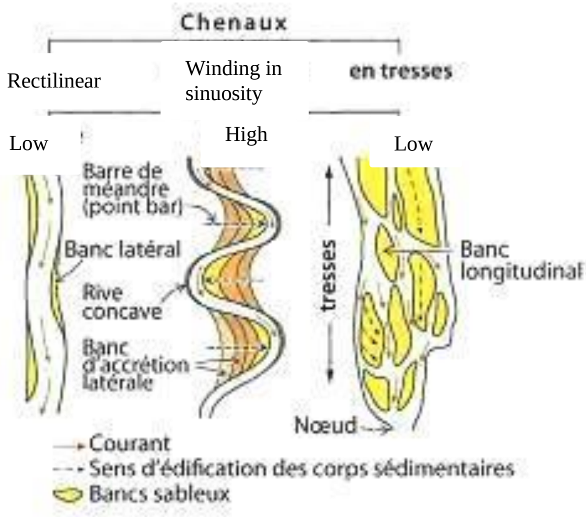


Torrential valley **Fig. 15:** General organization of an alluvial fan (numbers indicate successive lobes)

A-4. The fluvial networks

The type of network mainly depends on the slope, the transported load, and the stability of the banks (with the stabilizing role of vegetation). There are four types of channels: straight channel, meandering channel, braided channel, and anastomosed channel. The channel layout changes frequently during successive and sudden floods, with deposits usually being coarse.

The braided channel fluvial network is a watercourse with numerous unstable channels, forming divisions or connections between these arms, known as anastomoses. These different arms create a complex and rapidly shifting network, taking on a form reminiscent of a braid, hence the name.



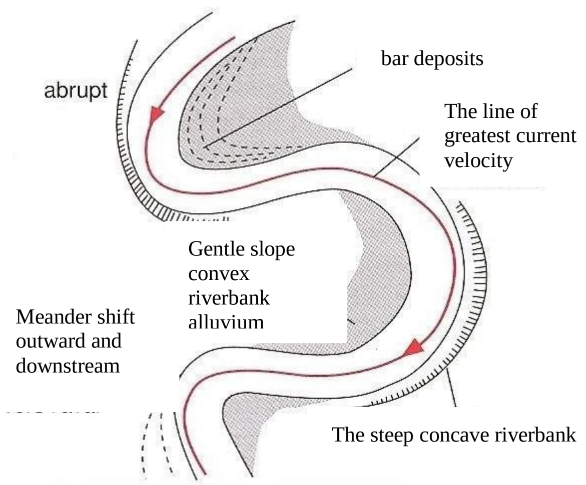
In a meandering fluvial network, the watercourses are slow and mainly transport fine sediments. Deposits cling to the convex banks of the channels, forming meander bars. The bar grows laterally as the meander migrates (lateral accretion).

The network is typically braided upstream and meandering downstream. Straight networks are rare. Anastomosed networks are observed in subsiding areas with a humid climate. Braided and meandering networks are the most common.

Fig. 17: Network of anastomosed channels

Fig. 18: Meandering network

Fig. 19: Elements of a meander



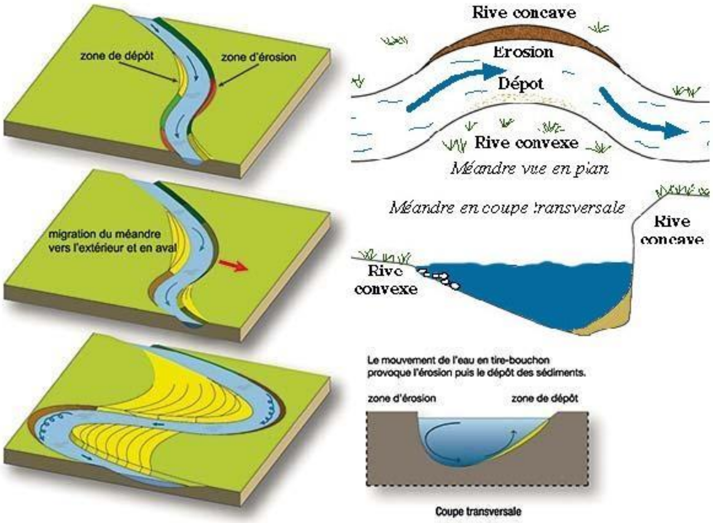


Fig.20: Cross-section and top view of a meander

A-5. The lacustrine environment

A lake is a permanent body of water enclosed within the continent, typically consisting of freshwater. Lakes exhibit a wide range in size, from shallow marshes to vast inland seas like the Great Lakes of America. Salinity levels also vary significantly, with highly saline bodies such as the Great Salt Lake and the Dead Sea considered lakes, alongside less saline ones like the Caspian Sea and the Black Sea in comparison to seawater. The origins of small lakes are diverse, including coastal plain lagoons (e.g., the Thau pond), abandoned meanders in alluvial plains, deltaic plain lakes (e.g., the Vaccarès pond), glacial overdeepening, volcanic crater lakes, limestone or karstic dissolution lakes influenced by meteoric waters, among others. Larger lakes typically have tectonic origins, such as fault lakes resulting from graben collapses in the African Great Rift Valley, the Dead Sea, etc.

Due to their isolation, lake characteristics vary according to climate, river inflows, geological settings, shoreline vegetation, and biological activity within the lake. Confinement is a common occurrence, leading to bottom anoxia. Accumulation of organic matter in substantial quantities results in the formation of sapropel (black mud) or lignite (woody debris). In arid climates, intense evaporation leads to the precipitation of salts on the shores (e.g., gypsum, halite, etc.).

Fig.21: Aerial view of Lake Geneva

The materials brought by rivers deposit in a lake in a somewhat theoretical concentric zoning that depends on hydrodynamics: pebbles along the shores, sands in the peripheral areas affected by waves, and mud in the deeper, calmer center. In reality, the distribution of materials depends on the location of deltas within the lake.

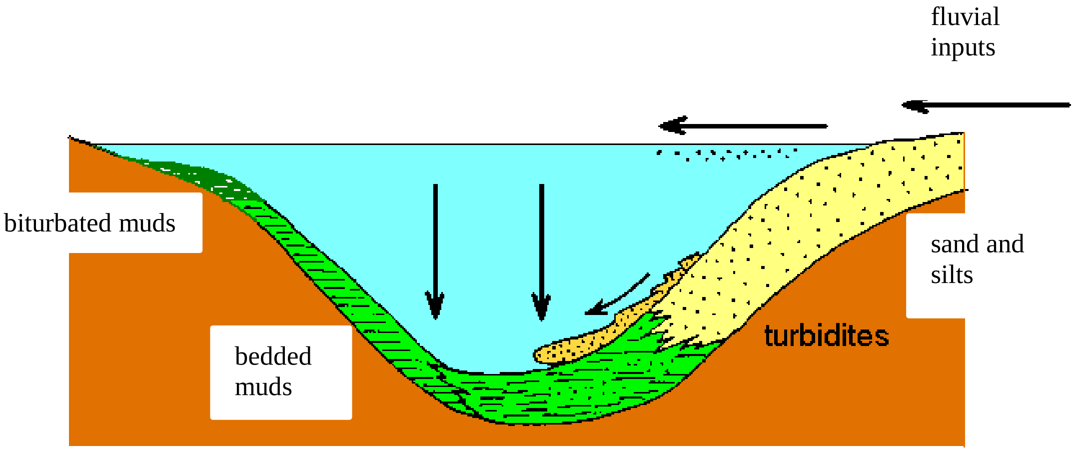


Fig.23: Schematic cross-section in an oligotrophic lake

A-6. Mixed Environments

Mixed environments represent intermediary zones between the landmass and the ocean, where estuaries and deltas act as outlets for continental rivers and streams into the sea. Here, freshwater blends with saltwater, transporting substantial quantities of sediment that disperse in the sea in diverse manners.

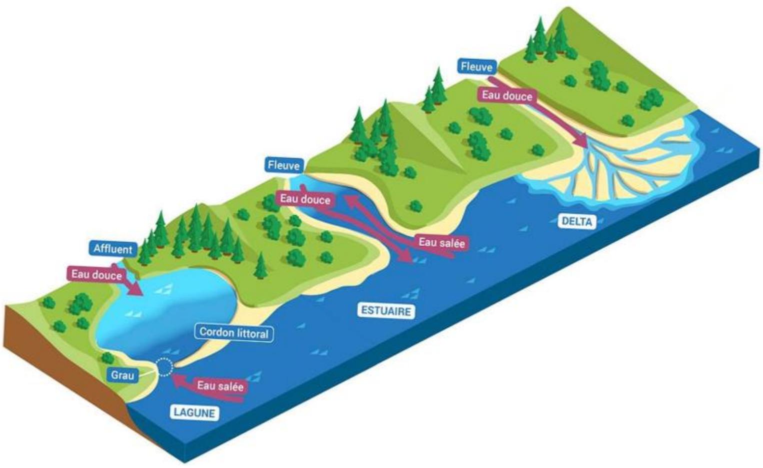


Figure 14: Block diagram illustrating intermediate environments (delta, estuary, and lagoon)

An estuary is the section of a river's mouth where the effect of the sea or ocean into which it flows is perceptible. The predominant influence is from the sea, and the primary deposit found in estuaries is mud.

A delta is a low sedimentary structure that experiences a significant accumulation of sediments due to the decrease in current velocity (large input from the river and minimal action from the sea),

like the Nile Delta flowing into the Mediterranean Sea.

Deposits consist of sequences of sandy bodies topped with clays.

Figure 15: Nile Delta (viewed from space)

The delta is categorized into 3 zones:

- The deltaic plain: above water level, primarily consisting of clay rich in plant-derived organic material, intricately dissected by numerous small channels formed by the river.
- The deltaic front: the juncture where freshwater and saltwater converge, witnessing sediment deposition of very fine sand and clay (silt).
- The pro-delta or delta's onset, marking the transition to the continental shelf, where clays and silts are deposited.

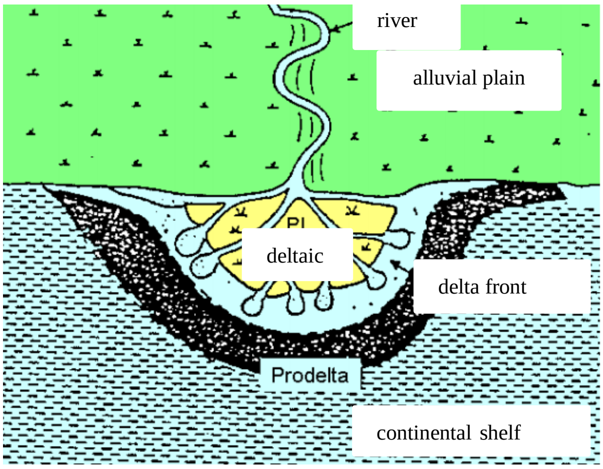


Fig.16: Components of a delta

A lagoon is a typically shallow environment where seawater is temporarily trapped without a direct water source due to a natural barrier. Often composed of fine sand, this barrier naturally evolves and is vulnerable to sea assaults (storms, tsunamis). This coastal water body is in varying and often permanent connection with the sea and sometimes one or more rivers (estuarine lagoons). Communication with the marine environment can occur through one or more openings, sometimes referred to as "inlets", which can be either permanent or temporary. There may also be an indirect connection via underground water tables crossing the permeable sediment barrier.

The lagoon's salinity levels vary depending on its exchanges with the sea and the contributions from the watershed, which fluctuate with the seasons. This saline water is often driven by tidal currents.

Consequently, a lagoon develops salinity gradients, with one end at the river mouths and the other at the sea inlet channels, potentially generating density currents. As a result, the sediments in a lagoon consist of various materials, including detrital elements brought by rivers and those formed by the precipitation of dissolved salts through evaporation and other inputs.

The sediment deposits in a lagoon are primarily influenced by thermal factors linked to the climate, except at the mouths of tributaries and tidal channels where currents may play a relative role. The lagoon area is therefore the preferred domain for evaporites.

Figure 17: Glenrock Lagoon in Newcastle, Australia **Fig.18:** Garabogaz-Gol Lagoon (Turkmenistan)

The biological composition, significantly impoverished compared to that of the nearby sea, faithfully reflects the physico-chemical characteristics of the lagoon environment. Depending on the situation, the fauna may be marine or, more commonly, typical of brackish waters. When salinity levels become too high, organisms disappear, and the sediments become azoic.

A-7. marine environment

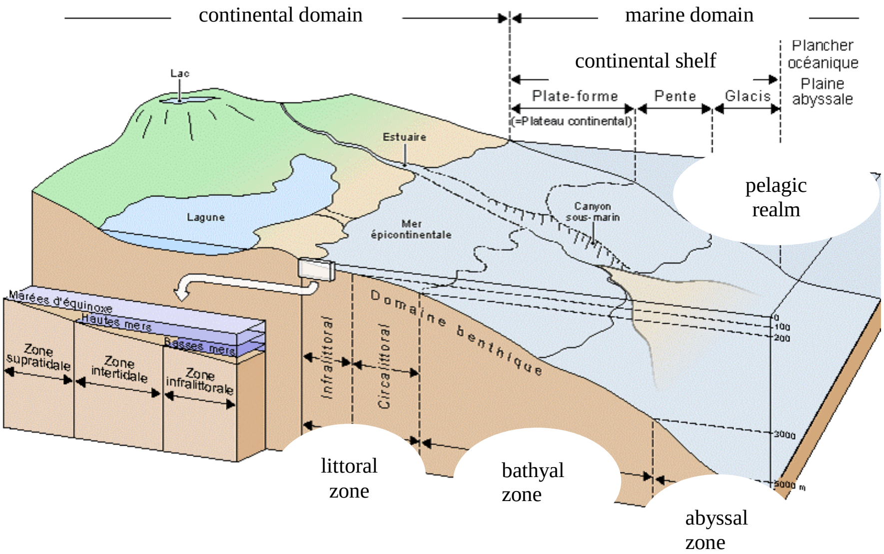


Fig. 19: Main domains of the marine environment

These environments are distinguished from one another based on the morphological characteristics of the seabeds and the correlated variations in water depth and distance from the shore. Sedimentation near the shore is heavily dependent on wave energy, while elsewhere it is influenced by marine currents.

At the edge of most current land masses, the continent extends beyond the shore into a submarine plateau known as the continental shelf. This is followed by a steep drop-off, which forms the edge of the continental shelf, leading to a sloping surface called the continental slope, collectively known as the continental margin. Beyond this, connecting to the continental slope through a kind of gently sloping area called the continental rise, lie the oceanic depths. These various regions of the ocean floor can be observed, for example, along the Atlantic coast of France and off the coast of North America.

The littoral zone is an area that receives sediments mainly of continental origin, especially detrital sediments. The movement of the waves helps to mix these sediments with debris from shells of coastal living organisms.

The continental shelf is a gradually sloping platform that extends towards the sea. It can span up to 80 km and reach depths of 200 meters. This area is characterized by the strong water dynamics influenced by tides, waves, and storms. Sediments are subjected to erosion, transport, and deposition when the current speeds decrease, leading to the presence of features like knobby sandstone and cross-stratified sandstone. In the calm deep waters, the breakdown and sedimentation of organic elements occur, precursor to special rocks like phosphates.

The continental slope, with its approximately 5° slope, allows sediment sliding, creating turbidity currents. These sediments eventually settle in the abyssal zone, forming plankton-rich turbidites, especially composed of calcareous organisms like foraminifera.

The deep sea, with its lack of strong currents, facilitates the deposition of the finest particles and suspended elements in the water, such as clays, siliceous plankton skeletons like radiolarians, as well as iron and manganese hydroxide salts, resulting in the formation of red clays.