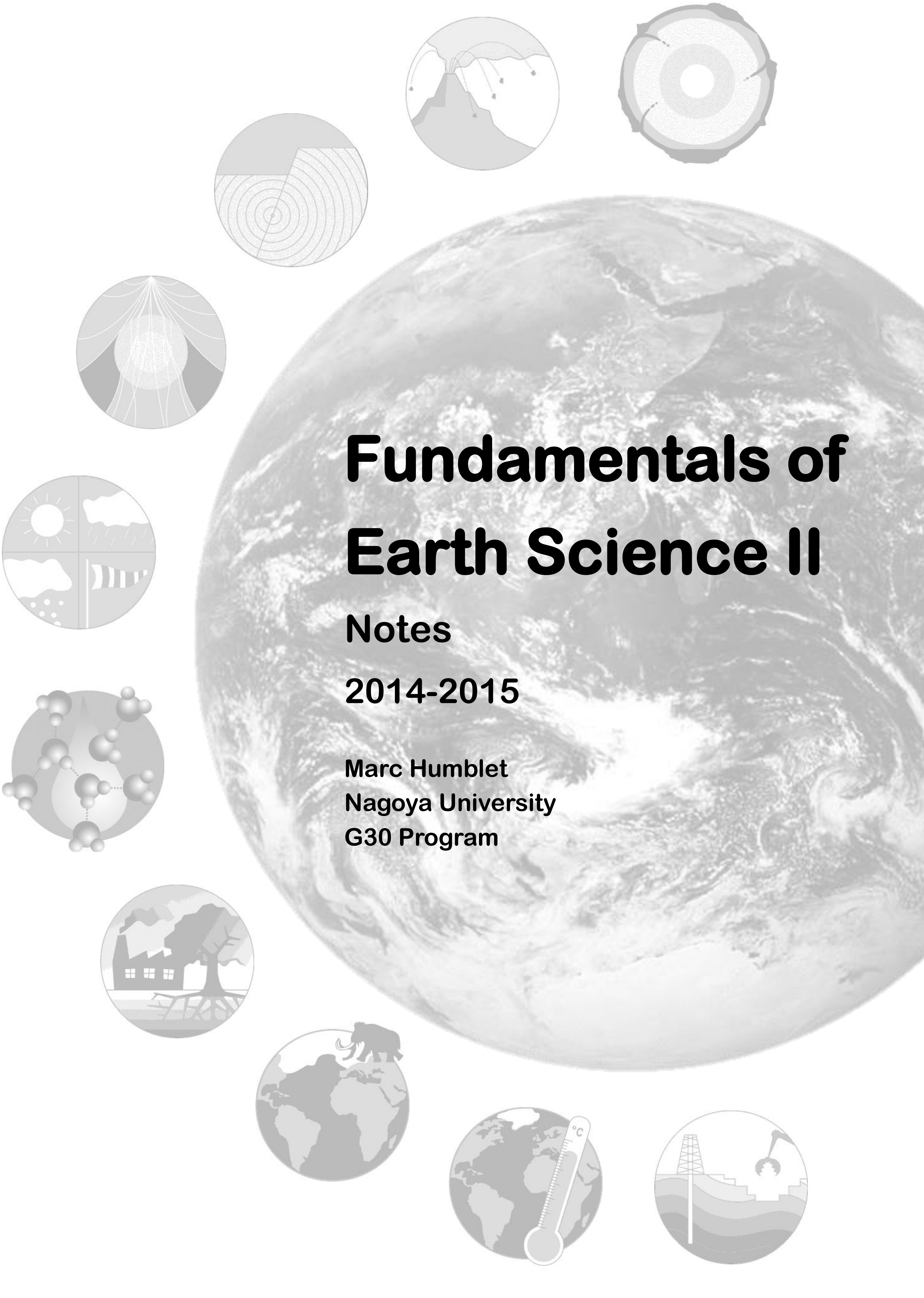

Fundamentals of Earth Science II

Notes

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1. The Earth system

1.1. introduction

The Earth system can be divided into three subsystems: (1) the ***climate system***, (2) the ***plate tectonic system***, and (3) the ***geodynamo system***. The first is driven by an external source of heat (the Sun), whereas the second and the third are driven by an internal source of heat —heat accumulated at the beginning of Earth's history and generated by the decay of radioactive elements.

(1) The ***climate system*** includes all the components of the Earth system whose interactions control Earth's climate: the ***atmosphere*** (layer of gas surrounding the Earth), the ***cryosphere*** (surface ice and snow, e.g. polar ice caps, mountain glaciers, permafrost), the ***hydrosphere*** (liquid surface water, including groundwater), the ***lithosphere*** (rigid rocky outer layer of Earth), and the ***biosphere*** (all living things). Recent global warming calls to our attention the role of human activities on climate change. So much so that it is adequate to define another component of the climate system: the ***anthroposphere*** (sum of all human activities influencing the environment).

(2) The ***plate tectonic system*** involves all the components of the Earth which control the movement of continents, the formation of mountains and ocean basins, and events such as volcanic eruptions and earthquakes. These components are the ***lithosphere***, ***asthenosphere***, and the deep ***mantle***. These terms will be described in the following paragraphs. As we will see later, characteristics of the lithosphere such as the position of continents, the presence of mountain chains or the shape of ocean basins influence the climate system.

(3) The ***geodynamo system*** produces and maintains Earth's magnetic field and involves movements of matter within the Earth's core.

Figure 1: Model of the Earth system and its three subsystems (i.e. climate, plate tectonics, and geodynamo).

1.2. Earth structure and plate tectonics

The exact age of the Earth cannot be determined directly but a good approximation can be obtained by measuring the age of meteorites. Meteorites are rocks falling from space on the surface of the Earth. They originate from the collision and fragmentation of larger bodies. These larger bodies —asteroids— are leftovers from the process of planetary accretion. In other words, they did not aggregate with other asteroids to form planets. The age of the Earth derived from meteorites is 4.6 billion years.

During the first billion years, the solar system was not yet cleared of the majority of its asteroids. This meant a higher probability of collision among asteroids and planets, and therefore a higher number of meteorites falling on the Earth. This period of Earth's history is referred to as the ***Heavy Bombardment***. Heat released by the collision between Earth and large meteorites was enough to melt Earth's surface. Another important source of heat was provided by the decay of radioactive elements contained inside the Earth. Consequently, our planet was occasionally in a molten ("soft") state, during which elements could migrate freely and matter redistributed according to its density. Heavier elements (Fe, Ni) migrated toward the center of the Earth, whereas lighter elements remained at the surface (Mg, Al, K). This process, by which Earth became a layered planet, is called ***gravitational differentiation***.

Earth is composed of 3 layers: the ***crust***, the ***mantle*** and the ***core*** (Fig. 2A). Each layer has a distinct chemical composition. Temperature and pressure increase toward the center. Only the outer core is in a liquid state. The inner core is solid because the pressure is extremely high and "forces" matter into a solid state despite the very high temperature.

Figure 2: Schematic cross section of the Earth. (A) Earth's main layers and (B) convection movements in the mantle (dark green = colder material sinking, yellow-orange = hotter material rising).

Most of the Earth's volume consists of the solid mantle. Solid does not mean totally rigid in this case. Slow, plastic deformations can take place, allowing matter to move. The rock "flow" inside the mantle is driven by differences of temperature/density. Hotter material rises toward the surface whereas cooler material sinks (Fig. 2B). This vertical motion is called ***convection*** (like convection taking place in a hot miso soup in which colder, denser soup in contact with the bowl sinks and hotter, lighter soup rises, creating convective cells).

Rocks of the crust and mantle are composed primarily of minerals of the ***silicate*** family. The basic structural unit of silicate minerals is $[SiO_4]^{4-}$, in which each oxygen atom shares one electron with the silicon atom. The crust is divided into an oceanic and continental crust. The former is thinner (up to 7 km thick), heavier and enriched in Fe and Mg[*], and lies beneath the ocean floor. The latter is thicker (up to 40 km thick), lighter and enriched in Al, K, and Na, and rises above sea level to form the continents (Fig. 3).

Figure 3: Thickness, composition and density of Earth's layers. Note the distinction between crust and mantle, and between lithosphere and asthenosphere.

The crust and the uppermost part of the mantle form a layer called the ***lithosphere*** (Figs. 3). The lithosphere beneath the oceans (oceanic lithosphere) has an average thickness of 70 km. Beneath continents (continental lithosphere), it can be more than 200 km thick. Directly below the lithosphere lies the ***asthenosphere***, a portion of the mantle a few hundred kilometers thick. The

[*] Note that the mantle is also rich in iron and magnesium. The chemical composition of the oceanic crust is very similar to that of the mantle because the oceanic crust is derived from the partial melting of mantle rock (see page 3).

physical properties of the two layers are very different. The lithosphere is rigid and brittle. The asthenosphere is weak and ductile.

The lithosphere is broken into plates called ***tectonic plates*** (Figs. 4 & 5). Tectonic plates move relative to one another and slide over the asthenosphere. The motion of plates is driven by convective movements inside the mantle. Three major types of plate boundary can be identified: (1) divergent, (2) convergent, and (3) transform-fault.

Figure 4: The main tectonic plates. A, B and C are the 3 types of plate boundary: (A) divergent boundary, (B) convergent boundary, and (C) transform-fault boundary (note that only major transform faults are indicated on this figure).

Figure 5: Global topographic map. Note the correspondence between the location of plate boundaries and Earth's topography. The deepest places on Earth consist of oceanic trenches along subduction zones and submerged mountain chains correspond to mid-ocean ridges (see text for explanations). Source: NOAA.

(1) ***Divergent boundaries***: two plates are pulled apart where hot mantle material rises and a large valley (***rift***) forms in between. Mantle rock rising toward the surface partially melts[*]. Some of the molten rock (magma) solidifies before reaching the surface; some reaches the surface forming a volcanic chain in the middle of the valley. This process is responsible for the formation of new oceanic lithosphere along plate boundaries called ***mid-ocean ridges*** (e.g. the Mid-Atlantic Ridge, Figs. 5 & 6A). The divergence of oceanic plates at mid-ocean ridges and the production of new oceanic lithosphere is a process referred to as ***sea-floor spreading*** (average spreading rate of mid-ocean ridges = 50 mm/yr).
The other type of divergent boundary is the ***continental rift*** where continental lithosphere is being pulled apart (e.g. the East African Rift, Fig. 6B), eventually leading to the formation of a new ocean basin.

(2) ***Convergent boundaries***: since the Earth's surface does not increase in time, the continuous production of oceanic lithosphere at mid-ocean ridges means that it must be destroyed somewhere else. This happens at ***subduction zones*** (Figs. 6C). At subduction zones, two plates converge (two oceanic plates or one oceanic and one continental plate). The heavier oceanic plate slips under the other one[**], and sinks deeper into the mantle where it is "recycled". The subducting oceanic plate sinks because it is colder and denser than the surrounding mantle. Subduction zones are characterized by a deep ***oceanic trench*** on the subducting plate side (e.g. the Mariana Trench) and by a ***mountain chain*** on the overriding

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[*] In this setting, the partial melting of mantle rock is due to the decrease in pressure during its ascension toward the Earth's surface (***decompression melting***). The melting temperature of a rock is lowered when pressure decreases.
[**] In case of a convergence between two oceanic plates, the older, cooler, hence denser plate subducts beneath the other one.

plate side (e.g. the Andes, Fig. 5). Subduction zones are also associated with volcanic activity[***] (e.g. Mount Fuji).

The other type of convergent boundary involves the collision between two continents (***continental collision***, Fig. 6D). In this case, the converging plates are both continental, hence light compared to the mantle, and no subduction can take place. Instead, a large mountain chain builds up where the two continents collide (e.g. the Himalayas, Fig. 5).

(3) ***Transform faults***: along transform-fault boundaries, two plates slip past one another (Figs. 6E-F). They most commonly offset mid-ocean ridges but they can also be found on land (e.g. San Andreas Fault).

Earthquakes occurs along all plate boundaries, whereas the majority of ***volcanoes*** are located along subduction zones and mid-ocean ridges.

Figure 6: Schematic cross section of tectonic plate boundaries. (*) Oceanic transform fault viewed from above (TF = transform fault, MOR = mid-ocean ridge). Note the motion of plates in opposite directions between the two segments of mid-ocean ridge, which triggers earthquakes.

1.3. Minerals: building blocks of rocks

Rocks are usually composed of an aggregate of different minerals although they can be composed of a single mineral too. Minerals are solid inorganic substances with atoms arranged in a regular, repeating pattern (***crystal lattice***). They have a specific chemical formula (e.g. NaCl, $CaCO_3$; Fig. 7). The external shape of mineral crystals reflects the internal order of atoms.

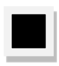

Figure 7: Crystal structure of halite (table salt). The green cube represents the repeating structural unit of the crystal lattice of halite (unit cell). Note the cubic shape of crystals of halite. Source (photograph): Understanding Earth 6th edition.

The structure of a mineral (the way atoms are stacked) depends on its chemical composition as well as the conditions of temperature and pressure during crystallization (Fig. 8).

Figure 8: Two minerals with the same chemical composition (carbon) but with different crystal structures. (A) Diamond formed at high temperature and pressure, and (B) graphite formed at much lower temperature and pressure. Source: modified from Understanding Earth 6th edition.

[***] Rock melting at subduction zones is facilitated by the presence of H_2O contained in sedimentary rock pores and clay minerals. The effect of water is to lower the melting point of rocks because water helps break chemical bonds (***water-induced melting***).

Three basic processes are responsible for the formation of minerals:

(1) Minerals form in saturated fluids when dissolved ions have reached their solubility threshold (e.g. precipitation of salts as water evaporates in a saline lake, Fig. 9A).
(2) Minerals form by solidification (liquid-solid transition) or deposition (gas-solid transition) (e.g. crystallization of a cooling magma, Fig. 9B)
(3) Minerals form by biological processes (***biomineralization***, e.g. corals, mollusk shells; Fig. 9C)

Figure 9: Illustration of the three processes of mineral formation. (A) Salts deposited at the bottom of a saline lake in southeastern Tunisia (scale bar = 2.5 km, source: Nasa Earth Observatory website), (B) rock composed of an aggregate of minerals crystallized during the slow cooling of a magma within the Earth's crust (granite, scale bar = 1 cm), and (C) coral skeleton composed of calcium carbonate (the specimen is 11 cm across).

1.4. Rocks and the rock cycle

The study of rocks is important because their mineral and chemical compositions and fossil contents can be used to reconstruct Earth's history and understand how life evolved. Rocks contain groundwater used for agriculture, public consumption and industrial purposes. They also contain mineral resources, such as gas, oil, coal and ore minerals, which are important from an economic and technological viewpoint. Their study is also crucial to solve environmental problems, such as the storage of radioactive substances and carbon dioxide, and the underground diffusion of pollutants.

Rocks can be classified into three families:

(1) ***Igneous rocks*** form by solidification of a cooling magma (molten rock). The size of mineral grains depends on the cooling rate (Figs. 10). Fine-grained igneous rocks are those which form near or at the surface of the Earth's crust when magma cools down rapidly (***extrusive igneous rocks***). Volcanoes form where magma reaches the surface (more or less violently). If the cooling rate is extremely high (e.g. magma in contact with air or water), crystals may not even have time to form, leading to the formation of a rock called ***volcanic glass***. Coarse-grained igneous rocks are those which form deep within the Earth's crust (***intrusive igneous rocks***). They result from the slow cooling of magma within the lithosphere (***magmatic intrusion***).

Figure 10: Examples of igneous rocks (hand-size specimens, scale bar = 2 cm, and photomicrographs taken in cross-polarized light, scale bar = 1 mm). (A) Intrusive igneous rock (granite) with large crystals formed in a slowly cooling magma, (B) extrusive igneous rock (basalt) with tiny crystals formed in a rapidly cooling magma, and (C) extrusive igneous rock (basalt) with tiny crystals mixed with volcanic glass (in black on the photomicrograph) resulting from a very high cooling rate (note the shiny surface of the sample). Source: Imperial College Rock Library.

(2) **Sedimentary rocks** form by accumulation and subsequent ***lithification***[*] of ***sediments*** (fragments of preexisting rocks or elements of biological origin, Figs 11A-C) or by precipitation of minerals from an aqueous solution (Fig. 11D). Sedimentary rocks forms in depressions of the Earth's crust where sediments can accumulate. Most originates in the largest depressions: the ocean basins. Sediments can be fragments of igneous, sedimentary or metamorphic rocks, or fragments of minerals (Fig. 11A). Sediments can also be of biological origin, such as mollusc shells, coral skeletons, bones and plant remains (Fig. 11 B-C). Other sedimentary rocks form by precipitation of minerals in evaporating lakes or embayments and are called ***evaporites*** (Fig. 11D).

Figure 11: Examples of sedimentary rocks (scale bar = 1 cm). (A) Conglomerate composed of rounded rock fragments (source: Imperial College Rock Library), (B) limestone composed of fragments of shells made of calcite ($CaCO_3$) (black arrows indicate shells of single-celled organisms called foraminifera), (C) coal (rock derived from an accumulation of plant debris, source: USGS) , and (D) evaporite composed of layers of anhydrite ($CaSO_4$) (source: Garcia-Veigas et al., 2013).

(3) ***Metamorphic rocks*** form by transformation of the chemical and/or mineralogical composition and/or texture[**] of a preexisting rock in a ***solid state*** due to changing conditions of temperature and/or pressure or due to interactions with hydrothermal fluids (Fig. 12). This process of rock transformation is called ***metamorphism***. Rocks caught in subduction zones or between two colliding continents are subject to tremendous changes in temperature and pressure causing rock metamorphism (***regional metamorphism***). In addition, wherever rocks are in contact with magma, these rocks, if not melted, are transformed (metamorphosed) —in this case "cooked"— by increasing temperature (***contact metamorphism***). Water heated in the vicinity of a magma circulates in the crust (***hydrothermal circulation***) and reacts with surrounding rocks, changing their chemical and mineralogical compositions (***metasomatism***). For instance, seawater which penetrates the oceanic crust through fractures near mid-ocean ridges is heated and leaches metals and sulfur from surrounding rocks. These elements precipitate as metal sulfides at underwater hot springs, forming large chimneys rising from the sea floor (***black smokers***).

Figure 12: Examples of metamorphic rocks (hand-size specimens, scale bar = 2 cm, and photomicrographs taken in cross-polarized light, scale bar = 1 mm). (A) gneiss showing an alternation of darker and lighter layers with different mineralogical compositions (layers are perpendicular to the direction of pressure —indicated by the arrows— affecting the rock during metamorphism), (B) schist with elongated minerals oriented at right angle to the direction of pressure (foliation), and (C) marble (metamorphosed sedimentary rock, usually limestone. Source: Imperial College Rock Library.

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[*] Lithification means solidification of a soft sediment (another word is ***induration***). This process occurs primarily by compaction (from the overlying sediment load) and precipitation of minerals in the space between sediment grains.

[**] The texture is determined by the size, shape and orientation of minerals.

The distribution of the different families of rocks is closely related to plate tectonics (Fig. 13). The formation of igneous rocks is linked to magma production and therefore to regions of the crust where rocks begin to melt. Rock melting takes place at mid-ocean ridges and subduction zones and at the root of large mountain chains related to continental collision. Most sedimentary rocks are produced in ocean basins whose formation is controlled by plate tectonics (i.e. the opening of ocean basins through sea-floor spreading). The distribution of metamorphic rocks too is of course closely linked to plate tectonics. Large variations of temperature and pressure affect rocks in convergence zones. Metasomatism related to hydrothermal circulation occurs in the vicinity of magmatic intrusions associated with plate boundaries.

Figure 13: Idealized cross section of the Earth's crust showing the regions of the crust where most igneous, sedimentary, and metamorphic rocks are produced. O-C SUBD. = ocean-continent subduction, HS = hot spot, O-O SUBD. = ocean-ocean subduction, MOR = mid-ocean ridges, CONT. COLL. = continental collision. Note that the hot spot location is not related to plate boundaries. Hot spots result from hot mantle material rising from great depths and producing volcanic activity at the surface (e.g. Hawaii).

Processes leading to the transformation of a rock of one family to a rock of another family can be described as a cycle called the ***rock cycle*** (Fig. 14). The rock cycle illustrates how each rock families can evolve from one another. The transformation of one rock family to another is controlled by plate tectonics and climate.

The control of plate tectonics on rock formation is clear when one looks at where the different types of rocks are being produced (see Fig. 13 and related text). Climate controls the production rate of sediments because the erosion of mountains depends on the amount of precipitation and other climatic factors (e.g. temperature). Hence, climate controls the production of sedimentary rocks. Plate tectonics also control the production rate of sediments because plate tectonics controls the formation of mountains from which sediments are derived. Moreover, plate tectonics influences the climate system through processes which will be discussed later in this course.

Figure 14: The rock cycle. Each arrow represents a transformation process from one rock family to another. Each transformation process is represented by a specific color. Note that metamorphic rocks themselves can also be metamorphosed.

2. Volcanism

2.1. Volcanoes and plate tectonics

Volcanism is the eruption of magma at the surface of a planet. Rocks produced by volcanic processes are called volcanic rocks. The most prominent feature of volcanism is the ***volcano***, an accumulation of volcanic material (e.g. molten rock, rock fragments, and gas) ejected from a central chimney or ***vent***.

First let's have a look at the distribution of volcanoes around the world and their relationship with plate tectonics (Fig. 15).

Figure 15: Global distribution of volcanoes and tectonic plates. Note that this map does not show underwater volcanoes.

The map above tells us that there is a close link between the distribution of volcanoes and plate tectonics. The great majority of volcanoes are located along plate boundaries. Most volcanoes displayed on the map are related to subduction zones. But volcanoes are also found along divergent boundaries (see the East African Rift in Fig. 15). Since the map above does not show volcanoes located below sea surface, those related to mid-ocean ridges are not displayed. However, one should keep in mind that underwater volcanoes may account for as much as 80% of Earth's volcanoes!

Volcanism occurs wherever magma reaches Earth's surface. This happens along tectonic plate boundaries: mid-ocean ridges, continental rifts and subduction zones[*]. But there is a third geologic setting related to volcanism that is not located along a plate boundary: ***hot spots***. Examples of hot spots are Hawaii (see red dots in the middle of the Pacific Plate in Fig. 15, see also Fig. 16), Yellowstone (continental hot spot in western USA), and Iceland (which happens to be also on the Mid-Atlantic Ridge). The mechanism of formation of these hot spots is not very well understood. A commonly accepted hypothesis suggests that the source of hot spots comes from deep within the mantle, perhaps even from the mantle-core boundary, and that a hot spot forms when a narrow jet of mantle rock called ***mantle plume*** reaches the surface.

Figure 16: Relief map showing Hawaii and the Emperor Seamount Chain (left) and sketch of a seamount chain related to the presence of a hot spot beneath a moving oceanic plate (right). As the plate moves toward the subduction zone, a succession of volcanoes is formed above the fixed hot spot. Volcanoes are carried away by the plate (much like a conveyor belt) and progressively sink as they (and the plate underneath) cool down and contract. At the same time, ocean waves and currents erode and carry away some of the volcanic material.

[*] Not in regions of continental collision (e.g. Himalayas) because the magma solidifies before reaching the surface.

Knowing how the temperature of Earth's interior evolves with increasing depth (***geothermal gradient***) is crucial to understand why volcanic activity occurs in a particular geologic setting. The geothermal gradient in the Earth's crust is on average 30°C/km. Its value greatly depends on the tectonic setting. For example, it increases sharply beneath mid-ocean ridges where the crust is thin and magma is formed near the surface. When the geothermal gradient intersects the rock solidus[**], the rock begins to melt and a magma is produced (Fig. 17, see figure caption for further explanations).

Figure 17: Cross section of the oceanic crust and mantle showing three geologic settings associated with volcanism: mid-ocean ridge, hot spot, and subduction zone. The red curve delineates the geothermal gradient (temperature vs. depth) and the green curve delineates the solidus of mantle rock. Note that the melting temperature (solidus) decreases with decreasing depth/pressure. The first graph to the left shows the geotherm of a portion of the oceanic crust that is not related to volcanic activity ("normal" situation). The geotherm and the solidus do not intersect and rock does not melt. In the three geologic settings where volcanic activity occurs, the geotherm and the solidus intersect and rock begins to melt once temperature reaches the solidus. Source: modified from Wikipedia (author: Woodloper).

2.2. The volcanic "plumbing" system

Magma usually originates in the asthenosphere by the processes described in figure 17. The magma, less dense (hotter) than surrounding rocks, makes its way up through the lithospheric mantle and crust by moving along fractures and by melting surrounding rocks. Once magma reaches the surface, it is called ***lava***. A volcano is made of successive layers of lava and/or ***pyroclasts*** (see section 5.6.). The path taken by the magma from its source to the surface delineates different geologic structures described in the figure below (Fig. 18).

2.3. Interactions between volcanoes and other geosystems

Volcanoes interact with other geosystems (e.g. atmosphere, hydrosphere; Fig. 19). As we will see later, volcanic gases (mainly water vapor, but also CO_2, SO_2, and lesser N_2, H_2, CO, S, and Cl) released in the atmosphere may influence the climate. Volcanic activity also influences the biosphere. Catastrophic volcanic eruptions have been a plausible cause of mass extinctions in the geological past. Very large volcanic eruptions may cause short-term global cooling due to volcanic aerosols[*] remaining in the upper atmosphere and blocking the sunlight, and may be responsible for acid rains and long-term global warming due to massive release of water vapor and CO_2 in the atmosphere. Volcanic activity is also closely related to the circulation of groundwater heated in the vicinity of a magma (***hydrothermal circulation***), and therefore has an impact on the hydrosphere too. Hydrothermal water returns to the surface as ***hot springs***. Hot springs are sometimes

[**] The solidus of a rock is its melting temperature. Below the solidus the rock is completely solid. Melting begins when the temperature reaches the rock solidus.

[*] Tiny solid particles and liquid droplets (e.g. sulfuric acid and water)

characterized by violent, intermittent eruptions of hot water called ***geysers***.

Figure 18: Cross section of a portion of the lithosphere through which magma is rising toward the surface. Magma typically accumulates in the crust in large ***magma chambers*** (1). Pieces of surrounding rocks can be incorporated in the magma and change its chemical composition (2). The magma starts to crystallize as it cools, which changes the composition of the remaining melt (3). Some of the magma leaves the chamber through fractures and forms large sheets, either parallel to rock layers —***sill***— (4) or intersecting rock layers —***dyke***— (5). A volcano is composed of a ***central vent*** (6) which may branch off and form a ***side vent*** (7). Lava is ejected from the vents and form ***lava flows*** (8). Other materials which are ejected from vents are ***pyroclasts*** (ash, volcanic bombs) and gases (9).

Figure 19: Interactions between volcanoes and other geosystems: atmosphere, biosphere and hydrosphere (see text for explanations).

2.4. Types of lavas and volcanic explosivity

Magmas/lavas may have different chemical compositions. Obviously, the composition of the rock from which the magma is initially derived (***parent rock***) is important in determining the composition of the resulting melt. Temperature is also crucial in controlling magma composition because different minerals begin to melt at different temperatures (Fig. 20).

Magmas produced at mid-ocean ridges and oceanic hot spots are derived from the partial melting of asthenospheric mantle rock. They are enriched in Fe and Mg. Their composition is referred to as ***mafic*** (from **ma**gnesium and **f**erric, Fig. 20). If such magma cools down slowly within the oceanic crust, the resulting intrusive igneous rock is called a ***gabbro***. On the other hand, if it reaches the surface and cools down quickly, the resulting extrusive igneous rock is called ***basalt***. Gabbro and basalt have similar compositions but differ by their texture. The former is coarse-grained whereas the latter is fine-grained. Gabbro and basalt are the two most common rocks of the oceanic crust. Basaltic magma has a very low content of silica and is very hot. The ***viscosity*** of magma is proportional to its silica content* and inversely proportional to its temperature. Therefore the viscosity of basaltic magma is relatively low, i.e. relatively fluid (Fig. 20). Hawaii is an oceanic hot spot and its volcanoes are characterized by fast-flowing, low-viscosity lavas and non-explosive eruptions. Magma viscosity determines whether a volcanic eruption is potentially more or less explosive. Highly viscous magmas tend to remain stuck in the volcanic vent and act like a cork.

Another important factor determining the nature of volcanic eruptions is the concentration of ***volatile compounds*** (e.g. HO_2, CO_2) present in the magma. A volatile-rich, highly viscous magma is much more likely to give rise to an explosive eruption. A magma nearing the surface

* Silica molecules (SiO_2) tend to form chains which greatly increase the viscosity of the magma.

behaves like a soda drink when you unscrew the bottle cap. The volatiles contained in the magma form gas bubbles which expand due to the pressure drop. Decreasing the amount of volatiles dissolved in the magma make it also more viscous. A volatile-rich, highly viscous magma is likely to give rise to an explosive eruption due to the pressure that progressively builds up in the chimney and is suddenly released in the form of a violent explosion.

Magma may gain components by melting the surrounding lithospheric rocks during its ascension toward Earth's surface (***crustal contamination***). If a basaltic magma initially derived from the partial melting of asthenospheric mantle penetrates the continental crust, it can melt the surrounding rocks and absorb elements from continental crustal rocks. Typical igneous rocks of the continental crust are ***granite*** and its extrusive equivalent, ***rhyolite***. Their composition is referred to as ***felsic*** (from **fel**dspar and **si**lica). Their color is lighter than mafic igneous rocks because felsic minerals are lighter in color than mafic minerals (Fig. 20). Mafic and felsic rocks are the two end-members of a continuum of rock/magma compositions. A basaltic magma traveling inside the continental crust can therefore become relatively more felsic as it melts surrounding rocks. Felsic lavas are highly viscous because there are not as hot as mafic lavas and contains a lot more silica (Fig. 20). Related volcanic activity will therefore be more explosive. Magmas with felsic or intermediate compositions may be produced in the following geologic settings: continental hot spots (e.g. Yellowstone), continental rifts (e.g. East African Rift), and subduction zones (e.g. Japan). In continental hot spots and continental rifts, the magma initially derives from asthenospheric mantle rock but becomes more felsic as it penetrates the continental crust and melts the surrounding rocks. In subduction zones, the initial melt is derived from asthenospheric mantle rock, oceanic crust and sedimentary rocks of various compositions (more or less silica-rich). In the case of an ocean-continent subduction, the magma can potentially become more felsic as it penetrates the continental crust. Note that in all these geologic settings (continental hot spots, continental rifts, and subduction zones) basaltic (mafic) volcanism can also take place.

In subduction zones (continent-ocean, ocean-ocean), the magma is often enriched in volatiles such as water (filling the space between sediment grains or in hydrated minerals, e.g. clay minerals) and CO_2 (from marine carbonate sediments). If the magma is also rich in silica, then the risk of explosive eruption is high. On the contrary, the magma produced at oceanic hot spots is characterized by a low volatile content, a high temperature, and a basaltic composition (low silica content), and therefore the risk of violent explosion is low.

We already mentioned (see caption of Fig. 18) that crystallization in a magma chamber influences the chemical composition of the residual melt. The process by which magma composition changes during crystallization is called ***Crystal fractionation***. The order in which minerals crystallize in a cooling magma has been determined experimentally. The idealized succession of minerals formed when magma cools is referred to as the Bowen series (see related slide). As magma crystallizes, the residual melt becomes progressively more felsic. Therefore the nature of volcanism (i.e. more or less felsic, more or less explosive) also depends on the timing of extraction of the residual magma[*].

Mixing of different magmas is another way to change magma composition (***magma mixing***) and influence magma composition, hence the nature of volcanism.

From the text above, you can easily see why it is so important to understand the process of magma formation. Magma composition has a huge influence on the type of volcanic eruption it triggers and the nature of the risk it involves for human populations.

Figure 20: Classification model of igneous rocks. Igneous rocks are classified according to their mineralogical and chemical compositions, here expressed by the proportion of selected minerals (expressed as a % of rock volume) as a function of the silica content (expressed as a % of rock weight). The upper part of the figure shows images of rock samples. Note the difference of color between felsic (lighter) and mafic (darker) rocks. Source: modified from Understanding Earth 6th edition (rock images: Imperial College Rock Library, except granodiorite: US National Park Service website, and peridotite: Wikipedia).

2.5. Types of eruptions and volcanoes

Two categories of eruptions can be distinguished: ***central eruptions*** and ***fissure eruptions***. The former are associated with volcanoes in which lava and other volcanic material are ejected from a central vent. The latter are eruptions during which lava erupts from fissures, cracks in the Earth's crust.

The type of lava influences the type of volcano. For example, ***shield volcanoes*** like Hawaii are formed by basaltic lavas related to an oceanic hot spot (Fig. 21A). In this setting low-viscosity lava easily flows out of the vent and spreads over an extensive area to form a large volcano with low-angle slopes. In contrast, ***volcanic domes*** are formed by highly viscous rhyolitic lavas which tend to remain in the vent and produce a dome-shaped cap on top of the volcano (Fig. 21B). Volcanoes composed primarily of pyroclastic layers (ashes and other larger fragments of volcanic rocks) are called ***cinder-cone volcanoes*** (Fig. 21C). These volcanoes are typically related to basaltic volcanism because a large fountain of magma (low-viscosity magma) is needed to deposit abundant fine-grained pyroclasts which form a cone. Successions of lava flows and pyroclastic layers are characteristic of ***stratovolcanoes*** (Fig. 21D). When the roof of a magma chamber that has been partially emptied collapses, it forms a large depression called a ***caldera*** (Fig. 21E). Another geologic structure related to central eruptions is called a ***diatreme*** (Fig. 21F). Diatremes are only known from the geologic record. No active diatreme system exists today. They

[*] Note that magma viscosity is also influenced by the crystals it contains. A magma that has begun to crystallize before reaching the surface and yields a high crystal content is more viscous than magma with a low crystal content.

consist of broad vents filled with fragments of rocks and solidified magma, and result from large explosive eruptions of material initially found at great depths in the mantle. Diamonds originating from the mantle may be found in some diatremes (e.g. the Kimberley diamond mine in Australia).

Fissure eruptions (Fig. 21G) are mainly associated to rifting along divergent boundaries (e.g. Iceland). Large amounts of basaltic lava can be produced during these eruptions (***flood basalts***). Gigantic flood basalts are found in the geologic record and are evidence of past volcanic eruptions whose intensities were much greater than those of modern eruptions (e.g. Columbia River Basalts in USA, Deccan Plateau in India). It is probable that these large eruptions have had a huge impact on the biosphere and may even have caused mass extinctions (e.g. end-Permian mass extinction related to Siberian Flood Basalts or Siberian Traps). Large amounts of volcanic ashes may also accumulate during fissure eruptions and form thick layers of ***ash-flow deposits***, another prominent feature of fissure eruptions.

Figure 21: Types of eruptions and volcanoes, the magma compositions to which they are related, and typical examples of volcanic edifices. Examples cited are those of active volcanic systems except the Kimberley diatreme (approx. 80 million years old). Source: modified from Understanding Earth 6th edition.

2.6. Volcanic rocks

Volcanic rocks comprise two types of rocks: ***extrusive igneous rocks*** (EIRs) and ***pyroclastic deposits*** (accumulations of ***pyroclasts***). EIRs form by rapid cooling of magma/lava near or at Earth's surface. Pyroclastic deposits are pieces of EIR of various sizes, shapes, and textures ejected during volcanic eruptions.

The texture of EIRs depends on the cooling rate of magma/lava. Slow cooling produces larger crystals. A common texture seen in EIR is characterized by larger crystals "floating" in a matrix made of tiny crystals (***porphyritic texture***, Fig. 22A). This texture is formed when a magma starts cooling slowly in the crust —forming large crystals— and the remaining melt cools faster as the magma approaches Earth's surface —forming tiny crystals—. A uniformly fine-grained texture indicates a uniformly rapid cooling very near or at Earth's surface (Fig. 22B). Cooling may be so rapid that crystals have no time to form. In this case, the resulting rock is an amorphous (non-crystalline) ***volcanic glass*** (***obsidian***, Fig. 22C).
Fast-cooling EIRs often display a vesicular texture (spherical cavities in the rock, Fig. 22E). This texture results from the formation of gas bubbles in the magma when the pressure drops during volcanic eruptions and gas is released and expands (think about what happens when you open a bottle of soda). ***Pumice*** is an extremely vesicular volcanic glass typically formed during violent explosive volcanic eruptions (Fig. 22E). The density of pumice is so low that it can float on water.

Pyroclasts are produced either by rock fragmentation during volcanic explosions or by the solidification of pieces of lava ejected in the air. They are classified according to their size. The finest pyroclasts are ***volcanic ashes*** (<2 mm, Fig. 23A). When they accumulate and become solid, the resulting rock is called a ***volcanic tuff*** (Fig. 23B). The accumulation of larger, angular rock fragments is called a ***volcanic breccia*** (Fig. 23C). Pieces of lava ejected in the air that are 2-64 mm in diameter are called ***lapilli***. Pyroclasts larger than 64 mm are called ***volcanic bombs*** (Fig. 23D). They typically have the shape of a rugby ball and can reach several meters across! Another volcanic hazard is the ***pyroclastic flow***. A pyroclastic flow is a mixture of hot ashes and gases moving down the slope of a volcano at very high speed (Fig. 24).

Figure 22: Examples of extrusive igneous rocks: (A) porphyritic andesite with large, light-colored crystals “floating” in a dark-colored matrix made of tiny crystals (scale bar = 2 cm, source: Imperial College Rock Library), (B) basalt composed uniformly of tiny crystals invisible to the naked eye (scale bar = 2 cm, source: Imperial College Rock Library), (C) obsidian —note the shiny, glassy surface of the sample—(source: USGS), (D) vesicular basalt (source: Geological Survey of Ireland), and (E) pumice —usually light in color—(scale bar = 1 cm, source: Ersoy et al., 2010).

Figure 23: Examples of pyroclasts and pyroclastic deposits: (A) volcanic ashes and SEM image of one single ash particle (scale bar = 30 µm, source: USGS), (B) outcrop of volcanic tuff (source: USGS), (C) volcanic breccia composed of fragments of diverse volcanic rocks (source: Colorado Geological Survey), (D) small volcanic bombs (scale bar = 2 cm, source: USGS).

2.7. Hazards vs. benefits

The danger of volcanic eruptions comes less from lava flows —you see them coming from far away— or volcanic bombs —unlikely to hit you unless you are very unlucky or very close to the crater— than from ***pyroclastic flows***, ***lahars*** and ***lanslides*** (Fig. 24). ***Pyroclastic flows*** (defined in the previous paragraph) are rapid and sudden events which can quickly engulf inhabited valleys in an extremely hot cloud of dust and gas. ***Lahars*** are torrential mud flows made of a mixture of pyroclasts and water flowing down the slopes of volcanoes. They can form in various situations: lava melting glacier ice, pyroclastic flow melting snow or mixing with a river, or rain falling on volcanic ashes. Another deadly volcanic hazard is ***landslide***. The flanks of a volcano can become unstable during an eruption and a portion may collapse and slide downhill. A large landslide occurring underwater or a large chunk of volcano sliding into the sea or into a lake can trigger a ***tsunami***, which is another possible cause of fatalities related to volcanic eruptions.

Figure 24: Illustration of volcanic hazards. Photographs: lahar (left) and pyroclastic flow (right). Source: USGS.

Unlike earthquakes, it is sometimes possible to tell whether a volcano is likely to erupt soon. Examples of indicators of an upcoming eruption are seismic activity (changes in the intensity and frequency of earthquakes), ground deformation (swelling of the volcano), and gas emissions (gas released due to pressure drop within a magma approaching the surface). In ideal cases, these signals allow to predict eruptions several days before they actually occur (although eruptions can occur without obvious detectable precursory signs!).

Volcanic activity has also qualities that are advantageous for human populations. Nutrient-rich ***volcanic soils*** are fertile and good for agriculture (Fig. 25A). Therefore many people's lives depend on volcanoes. Volcanic rocks, gases and mineral deposits related to volcanic activity may be of great economic value. Volcanic rocks may be used as building stones. ***Minerals and chemicals*** produced by volcanoes can be used by the industry. For example, perlite (a kind of volcanic glass) can be used for insulating buildings or filtering beer; boric acid can be used as antiseptic and insecticide; sulfur can be used as a component of fertilizers, cellophane, bleaching agent… Another important resource related to volcanic activity is ***geothermal energy***. The heat from geothermal fields can be harvested to warm buildings and to produce steam that can be used to generate electricity (Fig. 25B). The production of electricity from geothermal energy is particularly relevant today given the necessity to reduce our dependence on fossil fuels and nuclear energy.

Figure 25: Illustrations of the relationship between volcanoes and human populations. (A) Satellite view of the Sakurajima near Kagoshima city －note the green flanks of the volcano with cultivated fields at the foot and a densely populated area nearby－(source: Google Earth), and (B) geothermal power plant (background) with hot springs (foreground) in Iceland (source: Wikipedia).

3. Earthquakes

3.1. Elastic rebound theory

Rocks at the edges of tectonic plates are subject to tremendous forces resulting in intense deformation. The force per unit area acting on a rock is called ***stress***. The three types of directional stress experienced by rocks are compressional, tensional, and shear stress (Fig. 26).

Figure 26: Types of stress and associated types of fault. Arrows indicate the direction of forces applied to the rocks.

If the stress is large enough, rocks undergo deformation, i.e. a change of shape and/or volume. The amount of deformation experienced by a rock is called ***strain***. The behavior of a rock in response to stress can be ***elastic***, ***brittle*** or ***ductile***. A rock behaves in an elastic manner when it recovers its original shape after the stress is removed. When the stress exceeds a value called the rock ***strength***, the rock experiences a permanent deformation. The deformation can be either brittle in which case the rock breaks, producing a fracture called a ***fault*** (Fig. 27A), or ductile (plastic) in which case the rock is deformed without breaking. An example of ductile deformation is a ***fold*** (Fig. 27B). Note that rocks may first be folded and then fractured as stress increases.

Figure 27: Examples of rock deformations. (A) Brittle deformation (faults in northwestern Australia, source: Google Earth: lat.: -18.0556, long.: 126.5107) and (B) ductile deformation (folds in the Appalachian Mountains, USA, source: Google Earth, lat.: 40.562241, long.: -76.716128).

In the upper part of the lithosphere, rocks tend to behave in a brittle manner and the rock strength increases with increasing pressure (depth). Ductile behavior is promoted by high temperature and high pressure, which is why rocks of the asthenosphere can flow in convective currents. Decreasing the strain rate also promotes ductile deformation (think about the behavior of modeling clay as you stretch it slowly or fast). Another factor influencing the behavior of rocks is their mineral composition (e.g. peridotite becomes ductile at a higher temperature than granite –remember the high melting temperature of olivine–).

Earthquakes are produced by the brittle deformation of rocks. They are confined to the cold rigid lithosphere, mostly the crust, where rocks behave in a brittle manner. Rocks of the asthenosphere under conditions of high temperature and high pressure display a ductile behavior.

Prior to a rupture event producing an earthquake, a body of rock experiences a certain amount of elastic deformation during which energy is stored (***strain energy***). Once the stress applied to the rocks exceeds the rock strength, rupture occurs along a fault plane and the rocks on both sides of the fault slip past each other and recover their original shape (= elastic rebound, Fig. 28). During the rupture event, strain energy is released in the form of frictional heat and ***seismic waves*** (see section 7.3. for more details about the nature of seismic waves). The slipping motion of the two blocks starts at a point called the ***focus*** of the earthquake. The ***epicenter*** of the earthquake is the point on the Earth's surface directly above the focus. After the rupture event, stress buildup resumes until the next earthquake occurs. This is called the ***elastic rebound theory*** which explains the recurrence of earthquakes along active faults and is illustrated by the saw-tooth shape of the stress vs. time plot in figure 28.

Figure 28: Elastic rebound theory. Between t_1 and t_2, strain energy accumulates and the rocks undergo elastic deformation until rock strength is reached. Between t_2 and t_3, the rocks break along a fault plane and energy is released as heat and seismic waves. After t_3, strain energy starts to build up again. The recurrence interval of earthquakes in this model is the time interval between t_1 and t_3. Source: Understanding Earth 6th edition.

In reality, the frequency of earthquakes is not as regular as figure 28 suggests. If it was, it would be much easier to predict their timing. Irregularity can be attributed to the following factors (Fig. 29):

1. incomplete stress release during earthquakes
2. change in stress intensity (stress partially released by movement along a nearby fault)
3. change in rock strength (e.g. the infiltration of fluids in faults may decrease frictional strength due to the lubrication effect of fluids)

Figure 29: Factors influencing the frequency of earthquakes: (1) incomplete stress release, (2) change in stress intensity, and (3) change in rock strength. Blue lines represent the release of stress during successive earthquakes. Source: adapted from Understanding Earth 6th edition.

3.2. Earthquakes and plate tectonics

Most earthquakes are distributed along active plate boundaries. The depth at which they occur (***focal depth***) varies according to the tectonic setting (Figs. 30 & 31).
Earthquakes occurring along divergent margins are shallow and associated with tensional forces (Fig. 31A). Shallow earthquakes also occur along transform faults. Transform faults associated with mid-ocean ridges produce earthquakes along the portion of the fault where the two plates move in opposite directions (Fig. 31B). The deepest earthquakes are related to convergent margins, particularly subduction zones (Fig. 31B). In this setting, the cold, brittle oceanic plate plunges into the asthenospheric mantle. The subducting plate remains brittle until it reaches a depth where temperature and pressure are too high for brittle deformation to occur. The deepest earthquakes are generated at around 700 km. Earthquakes associated with a continental collision may be deep as well but not as deep as those related to subduction zones. The maximum focal

depth recorded in the Himalayas is approximately 100 km. Below that depth, continental crustal rocks tend to deform in a ductile manner. Remember also that the brittle strength of crustal rocks increases with increasing pressure (depth).

Earthquakes may also occur far from plate boundaries (Fig. 30). These ***intraplate earthquakes*** are linked to hot spots or old faults related to ancient plate boundaries (e.g. the 2011 Virginia earthquake in the United States occurred in an area which used to be part of a collision zone between two ancient continents).

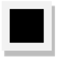

Figure 30: Global distribution of earthquakes. Here earthquakes are classified in three categories according to their focal depth: shallow (< or = 50 km, blue dots), deep (50-300 km, green dots), and very deep earthquakes (>300 km, red dots). Each dots represent a epicenter. Source: Understanding Earth 6th edition (M. Boettcher and T. Jordan).

Figure 31: Focal depth of earthquakes in different tectonic settings. The orange line delineates the lower limit of the seismogenic zone. (A) Mid-ocean ridge, (B) transform fault, (C) subduction zone, and (D) continental collision.

3.3. Seismic waves

During an earthquake, the rocks are affected by three types of movements corresponding to three distinct groups of seismic waves (Fig. 32):

1. ***P-waves*** (primary waves): compressional waves (push and pull motion, much like sound) traveling at 6 km/s in solid rocks and more slowly in liquids and gases.
2. ***S-waves*** (secondary waves): shear waves (sideways motion –at right angle to the direction of wave propagation–) traveling more slowly than P-waves and unable to propagate in liquids and gases.
3. ***Surface waves***[*]: confined to the surface of the Earth with two components:
    (1) the ***Rayleigh wave***: an elliptical motion decreasing with depth (similar to ocean waves).
    (2) the ***Love wave***: a lateral motion (sideways shaking).

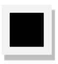

Figure 32: Seismic waves. (A) Primary wave, (B) secondary wave, and (C) surface wave (Love and Rayleigh waves). Source: adapted from USGS.

[*] Surface waves are responsible for most of the ground shaking during an earthquake.

When an earthquake occurs, P-waves and S-waves travel in all directions within the Earth. Surface waves travel exclusively at the surface of the Earth and are slower than P- and S-waves. S-waves travel more slowly than P-waves (Fig. 34). The speed of P- and S-waves depends on the physical properties of rocks. The harder the rock, the faster the wave. Therefore, wave speed depends on rock density –hence pressure, temperature and the rock mineral composition–. The relationship between the speed of seismic waves and rock density combined with the properties of wave refraction and reflection at the interface between Earth's layers can be used to reconstruct the structure of Earth's interior (see Chapter 7).

3.4. Earthquake characterization: localization, magnitude, and intensity

The measure of seismic waves amplitude and frequency is achieved by using an instrument called a ***seismograph*** (Fig. 33). A seismograph is built to record vertical and horizontal ground movements. The basic principle is very simple: a mass is loosely attached to a frame which is anchored to earth. The fixed mass is equipped with a pen which records the difference in motion between the mass and the frame on a revolving drum –the drum moves with the frame–. The mass must be heavy enough to remain at rest relative to the frame. A mass attached to a spring is used to record vertical ground movements (Fig. 33, right). A mass attached to a hinge is used to detect horizontal ground movements (N-S and E-W components, Fig. 33, left). The graphs obtained are representations of seismic waves and called ***seismograms*** (Fig. 34).

Figure 33: The seismograph. Instrument recording horizontal ground movements (left) and vertical ground movements (right). See text for explanations. Source: USGS.

Figure 34: Seismogram of a distant earthquake recorded in United Kingdom. Source: British Geological Survey.

How can we determine the location of an earthquake?

Since P-waves travel faster than S-waves, the greater the distance from an earthquake, the greater the difference between the arrival times of P-waves and S-waves (P-S intervals). The relationship between P-S intervals and the distance from the epicenter of an earthquake is predictable over thousands of kilometers. Seismologists have established travel-time curves that can be used to pinpoint the distance from the epicenter of an earthquake (Fig. 35). These travel-time curves are based on data from a great number of earthquakes and many seismographs around the world. Providing you have at least three seismographs in different geographic locations recording the same waves, you can locate precisely the epicenter of the earthquake. The focal depth can be determined based on further analyses of seismograms.

Figure 35: Travel-time curves of seismic waves and three seismograms from three stations recording seismic waves from the same earthquake. Matching these seismograms with the travel-time curves allows to determine the distance between each stations and the earthquake focus. Source: USGS.

How can we determine the size of an earthquake?

The size of an earthquake can be determined based on two different techniques:

1. Measure of the amount of energy released (***magnitude scales***).
2. Estimate of the destructive effects of ground shaking (***intensity scales***).

The first widely used magnitude scale was the ***Richter scale***. The Richter magnitude of an earthquake depends on the logarithm of the amplitude of the largest wave and the distance from the epicenter[*] –the time lag between the P-wave and S-wave– (Fig. 36). Magnitudes are expressed as whole numbers and decimal fractions (e.g. 3.5, 5.2, 6.4…). An increase of one unit of the Richter magnitude corresponds to a tenfold increase in wave amplitude (and 32 times more energy released[*])!

Figure 36: Graphic representation of the relationship between the Richter magnitude of an earthquake, the amplitude of the largest wave (in this example, 23 mm measured on the seismogram) and the difference in arrival time between the P wave and S wave (in this example, 24 s). Note that for a given distance from the epicenter, an increase in one unit of magnitude corresponds to a tenfold increase in the wave amplitude. Source: USGS.

A scale that is more widely used nowadays is the ***moment magnitude*** scale. The moment magnitude is a function of the seismic moment. The seismic moment is proportional to the area of faulting and the average fault offset[**]. Like the Richter magnitude, the moment magnitude can be calculated based on the analysis of seismograms. In the moment magnitude scale, an increase of one unit of magnitude corresponds to a tenfold increase in the area of faulting. The values of moment magnitudes are roughly similar to those of the Richter scale. For example, if an earthquake measures 5.5 on the Richter scale, its moment magnitude will be approximately 5.5 too.

The other way to evaluate the size of an earthquake is to establish a scale related to the destructiveness of earthquakes. This is the ***intensity scale***. An intensity scale that is commonly used today is the ***modified Mercalli scale*** (Fig. 37).

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[*] Richter's magnitude (M_L) = log A + 2.56 log D – 1.67, where A is the amplitude of ground motion (in μm) and D is the distance from the epicenter (in km) (source: British Geological Survey). Hence for a given value of D, each increment of 1 unit of magnitude M corresponds to a tenfold increase of the amplitude A.

[*] The energy released during an earthquake (E) is proportional to $10^{(1.5*M)}$. If E1 is the energy of an earthquake of magnitude M and E2 is the energy of an earthquake of magnitude M+1, E2/E1 = $10^{1.5}$ = 32 (source: British Geological Survey).

[**] Moment magnitude (M_W) = 2/3 log M_O – 6.06. The seismic moment (M_O) = μ ∗ rupture area ∗ slip length, where μ is the shear modulus of the crust ($3x10^{10}$ N/m^2). The shear modulus is the measure of the resistance of a material to shear strain (source: British Geological Survey).

Figure 37: Modified Mercalli scale compared with the Richter magnitude scale (source: Missouri Department of Natural Resources) and a modified Mercalli intensity map of the 1906 San Francisco earthquake (source: USGS, authors: John Boatwright and Howard Bundock).

In Japan, it is the intensity scale established by the Japan Meteorological agency that is preferentially used (the JMA or Shindo scale). It is analogous to the Mercalli scale but comprises only 7 levels of intensity instead of 12 (http://www.jma.go.jp/jma/en/Activities/inttable.html).

3.4. Fault mechanisms

In order to fully characterize an earthquake, we not only need to know its focus, origin time, magnitude and intensity, but we also need to describe the characteristics of the fault which triggered the earthquake or ***fault mechanism***: the direction of the ***fault strike***[*] (e.g. north-south, east-west), the angle and direction of the ***fault dip*** (e.g. 60° south, 75° east), and the ***type of fault*** involved (normal, reverse, or strike-slip –right lateral or left-lateral–, Fig. 38).

Figure 38: The three basic types of fault. Note that strike-slip faults can be either right-lateral or left-lateral. In the former case, one observer standing on one side of the fault sees the other side moving to the right (i.e. right-lateral). In the latter case, the observer sees it moving to the left (i.e. left-lateral).

For faults that can be observed at the surface of the Earth, the fault mechanism can be determined based on field observations. But for faults that are too deep to break the surface or faults at the bottom of the ocean, the fault mechanism must be deduced from the analysis of seismograms. Seismologists look at the ***first ground movement*** recorded by seismographs located at different places around the earthquake's epicenter. The first ground movement is determined on a seismogram by looking at the shape of the first P-wave. For example, on a seismogram recording horizontal ground movements, if the line of the seismogram goes up first, it means that the ground has moved away from the earthquake's epicenter (extension). If the line goes down first, it means the ground has moved closer to the earthquake's epicenter (compression) (Fig. 39). This allows the identification of two perpendicular planes, one of which is the fault plane. This method alone does not allow distinguishing the fault plane from the other one. Additional information is needed to determine which one is the fault plane.

[*] the fault strike is the intersection of the fault plane and a horizontal surface.

Figure 39: Schematic relationship between the first ground motion (horizontal in this example) recorded on seismograms and the fault mechanism. The figure shows a bird's eye view of the ground. The center of the large circle represents the earthquake's epicenter. Black dots are seismographs recording a first ground displacement away from the epicenter (line going up first). White dots are seismographs recording a first ground displacement toward the epicenter (line going down first). One of the two perpendicular lines delineating the four quadrants is the fault plane. Source: adapted from USGS.

3.5. Seismic hazards and risks

Seismic hazards are the natural phenomena related to earthquakes that may be potentially harmful to human populations. Seismic hazards directly caused by an earthquakes are called primary hazards: faulting and ground shaking. These may in turn cause secondary hazards, such as landslides, soil liquefaction, tsunami (see below for more details about these secondary hazards). ***Probabilistic seismic hazard maps*** display the probability that a given site will experience ground shaking exceeding a given value within a given period of time (see the end of this chapter for more information about earthquake prediction). Figure 40 intends to show the difficulty to accurately predict that a large earthquake will occur at a particular place within a certain period of time.

Figure 40: Probabilistic seismic hazard map of Japan showing the probability (0-100% chance it happens) of ground shaking of intensity "6-lower" or higher (on the JMA seismic intensity scale) during a 30-year period starting in January 2010. Note that the large 2011 Tohoku earthquake was not predicted by this model. Source: Robert J. Geller (2011) in Nature, v. 472, 407-409.

The ***Seismic risk*** is a measure of the damage expected in a given interval of time in a particular region. The seismic risk in a city with buildings engineered to resist large earthquakes will be much lower than in a city in which building codes are not adapted even if these two cities are exposed to the same seismic hazards. Seismic hazards must be determined prior to the assessment of seismic risks.

Besides ground shaking and faulting, three important secondary seismic hazards that can cause great damages are ***landslides***, ***soil liquefaction***, and ***tsunamis***. Landslides and soil liquefaction are due to ground instabilities resulting from ground shaking. The former occur when material on a slope becomes unstable and moves down the slope as a result of this instability (Fig. 41). Soil liquefaction refers to a layer of water-saturated sediment losing its cohesion and behaving as a liquid when shaken (Fig. 42). The shaking causes the grains to loose contact between each other, therefore allowing the sediment to flow like a liquid (Fig. 42A).

Figure 41: Landslides in El Salvador caused by a large earthquake in 2001 (left, source: USGS) and in Fukushima, Japan, caused by the 2011 Tohoku earthquake (right, source: Public Works Research Institute, Landslide Research Team, Erosion and Sediment Control Research Group).

Figure 42: Schematic representation of sediment liquefaction (left) and example of the damages caused by liquefaction during the 1964 Niigata earthquake, Japan (right, source: Japan National Committee on Earthquake Engineering, Proceedings of the 3rd World Conference in Earthquake Engineering, Volume III, pp s.78-s.105).

A tsunami represents another important secondary seismic hazard. The effect of a tsunami can be a lot more disastrous than ground shaking. During an earthquake, a tsunami forms when a portion of the seafloor moves vertically along a fault plane. The above water mass is displaced and creates a set of waves which propagate in all directions. Large tsunami can be generated at subduction zones where the overriding plate gets squeezed by the subducting plate and a considerable amount of stress accumulates. Stress is released when a portion of the crust springs seaward, raising the seafloor and pushing a large water mass up (Fig. 43). Tsunami waves travel very fast (up to 800 km/h) and have a very long wavelength (hundreds of kilometers between two successive waves). As they approach the shoreline, waves slow down by friction against the seafloor and their height increase (conversion of kinetic energy to potential energy).

Figure 43: Process of tsunami formation at subduction zone. (A) a portion of the overriding plate (1) is pushed landward by the subducting plate (2) and stress accumulates. (B) Stress is suddenly released as the portion of the overriding plate springs seaward along a thrust fault (low-angle reverse fault), generating an earthquake and pushing a water mass up. (C) The displacement of the water mass creates a set of tsunami waves at the surface which propagates in all directions.

Tsunami can travel over a considerable distance and can still be detected thousands of kilometers away from their origin (Fig. 44).

Figure 44: Simulation of the maximum wave amplitude related to the 2011 Tohoku tsunami based on data from buoys and sea-floor sensors. Source: NOAA.

When it comes to improving public safety in areas prone to earthquakes, there are two fundamental questions which have to be considered: (1) how can seismic risk be reduced? and (2) can earthquakes be predicted?

How can seismic risk be reduced?

- Hazard characterization
- Land-use policies
- Earthquake-proof engineering
- Emergency preparedness and response
- Warning systems (for earthquakes and tsunamis)

Can earthquakes be predicted?

With our current knowledge and technology, geologists cannot yet predict an earthquake early enough to evacuate the area. As we have seen earlier, probabilistic models exist but they make predictions over the long term (30 or 50 years) and sometimes fail to predict large earthquakes. In order to make predictions, geologists need to estimate the recurrence interval of earthquakes in a particular region. There are two ways to do it:

(1) One is to study historical data to infer the probability of future earthquakes. One can also go back further in time by looking at the geological record and dating earthquake-induced deformations of sediment layers or tsunami-lain sediments.

(2) Another way is to measure how much strain (deformation) accumulates along a fault every year (e.g. using GPS). Knowing when this fault slipped for the last time and how much strain was released, it is possible to estimate the timing of the next earthquake based on strain accumulation measurements. This kind of data is still relatively rare because it requires monitoring fault systems for a long time.

Earthquake prediction is complicated by the fact that faults are often not isolated but form complex networks. The release of stress in one fault segment may decrease or increase stress in another fault segment. It is therefore essential to understand how stress is distributed in these fault systems and how individual faults influence each other in order to improve the reliability of earthquake predictions. Geologists have been trying to use small earthquakes (foreshocks) occurring before larger seismic events to predict large earthquakes days or weeks before they occur. However, it is very difficult to distinguish foreshocks from the seismic background and the method is not yet reliable for most quake-prone areas.

4. A look at Earth’s interior using seismic waves

The bulk of the Earth’s interior cannot be studied directly. The deepest scientific boreholes are only a few kilometers deep. Tectonic forces sometimes bring deep crustal rocks to the Earth’s surface where they can be studied by geologists. Rocks from the mantle may also be transported to the surface during volcanic eruptions. Another source of information comes from lab experiments trying to recreate the very high pressure and temperature conditions occurring deep inside the Earth. Meteorites also provide information on the structure, composition and formation process of planetary bodies and, by analogy, the Earth as well.
Indirect methods of exploring the Earth’s interior are the domain of geophysics and use a variety of techniques related to:

- The behavior of seismic waves travelling inside the Earth
- Earth’s gravitational field
- Earth’s magnetic field
- The thermal properties of rocks

This chapter deals with the seismic waves as a tool to reconstruct the Earth’s interior.

4.1. Behavior of seismic waves traveling inside Earth

There are two types of seismic waves traveling through Earth’s interior: P-waves (compressional, “push-pull” motion) and S-waves (shear waves, “sideways” motion). As we saw in the previous chapter, P-waves are faster than S-waves. S-waves cannot propagate in liquids or gases. Seismic waves travel faster in denser materials. Hence, their speed generally increases with increasing depth (pressure) and is influenced by the chemical and mineral compositions of rocks. For example, P-waves travel more slowly in granite (6 km/s) than in basalt (7 km/s) or peridotite (8 km/s). Note that peridotite (upper mantle rock) is denser than basalt which is denser than granite.

P-waves and S-waves propagate in all directions. As they encounter the boundary between two layers, they can be either reflected or refracted. These reflected and refracted waves can be recorded by seismographs around the world. Knowing the distance from the epicenter and the arrival times of the different waves, information on the structure of the Earth’s interior can be obtained. The relationship between the angle of incidence and the angle of refraction is given by Snell’s Law of refraction (Fig. 45).

But let’s consider first a simple model with two layers –layer 1 and layer 2– and a seismic wave (P or S) generated by an earthquake near the surface (Fig. 46). The interface between these two layers is smooth and horizontal. In addition, we assume that layer 2 is denser than layer 1. Consequently, seismic waves travel faster in layer 2 than in layer 1 ($v_2 > v_1$).

Figure 45: Representation of Snell's law of refraction. If $V_1<V_2$, when the incident wave reaches the interface between layer 1 and 2 at the critical angle (i_c), the wave is refracted along the interface itself. The wave is reflected back into layer 1 for angles of incidence greater than the critical angle.

In this model, layer 1 could be the crust and layer 2 could be the mantle. We can distinguish 4 types of waves:

- ***Direct wave***: seismic wave traveling at the surface —therefore never reaching the interface between layer 1 and 2—(red lines in Fig. 46). This wave travels at speed v_1.
- ***Refracted wave***: seismic wave refracted at the interface between layer 1 and 2 when the angle of incidence is smaller than the critical angle (magenta lines in Fig. 46).
- ***Head wave***: seismic wave propagating along the interface between layer 1 and 2 at speed v2 (critical refraction) and refracted upward into layer 1 where its speed is equal to v1 (blue lines in Fig. 46).This wave is produced when the incident wave reaches the boundary at the critical angle. The angle at which the wave returns into layer 1 is equal to the critical angle.
- ***Reflected wave***: seismic wave reflected back into layer 1 when the angle of incidence is larger than the critical angle (green lines in Fig. 46).

Seismographs located at various distances from an earthquake's epicenter can be used to reconstruct the travel-time curves of the different waves (Fig. 46). They will detect the direct wave, the head wave, and the reflected wave at different times. Seismographs close to the epicenter will record the direct wave first. Beyond a certain distance, seismographs will start to "see" the head wave first. That is because the head wave travels along the interface between layer 1 and 2 at speed v_2, which is faster than the direct wave. The point on the travel-time curve from which the head wave is recorded prior to the direct wave depends on the thickness of layer 1 and on V_2. Knowing the slope of the travel-time curve of the head wave, we can determine v_2 (Fig. 46). By recording the arrival times of the first seismic wave reaching seismographs located at various distances from an earthquake's epicenter, it is therefore possible to calculate the depth of the boundary between layer 1 and 2. The reasoning for a 2-layer model is the same for models with more layers (Fig. 47).
In 1910, Andrija Mohorovičić (1857-1936), a Croatian seismologist, used travel-time curves of seismic waves to prove the existence of the crust-mantle boundary and to determine its depth. This boundary is called the Mohorovičić discontinuity in his honor (referred to as the "Moho").

Figure 46: Behavior of seismic waves (P- or S-waves) generated by an earthquake in an idealized 2-layer model. Dotted lines represent wavefronts and plain lines terminated by arrows represent the paths of seismic waves. Four types of waves are recognized (direct wave, reflected wave, head wave, and refracted wave). The arrival times measured by seismographs are used to construct travel-time curves (reflected wave in green, direct wave in red, and head wave in blue). In the present example, the five seismographs closest to the epicenter record the direct

wave first (A), whereas the others record the head wave first (B).

Figure 47: 3-layer model showing the path of a direct wave (red line) and two head waves (blue lines). Seismographs (black squares) are located at various distances from the earthquake's epicenter. The arrival times of the first wave reaching these seismographs are plotted against the distance from the epicenter. Three lines can be recognized. The first represents the direct wave (slope $1/v_1$). The second line is the head wave refracted along the interface between layer 1 and 2 (slope $1/v_2$). The third is the head wave refracted along the interface between layer 2 and 3 (slope $1/v_3$).

4.2. Reconstruction of Earth's structure based on seismic waves

Seismic waves generated by a large earthquake may travel through the entire Earth. The speed of seismic waves increases with increasing pressure. The speed of waves therefore generally increases gradually as seismic waves travel deeper into Earth's interior*. Since speed increases gradually, so does the angle of refraction (Snell's law, Fig. 45). Therefore, the trajectory of seismic waves traveling inside the Earth is not straight but curved (Fig. 48).

Figure 48: (A) multi-layer model representing the increase in wave speed with increasing depth. As a consequence, the wave is refracted at an angle that becomes progressively larger and therefore delineates a curved path. (B) Schematic representation of the curved path of a seismic wave traveling through the Earth.

Information on the structure of the Earth and the composition of Earth's layers can be inferred from travel-time curves of seismic waves. The method that was used for the discovery of the Moho (crust-mantle boundary) can be applied to other discontinuities of the Earth's interior. For example, the German seismologist Beno Gutenberg (1889-1960) identified for the first time seismic waves that had been reflected on the mantle-core boundary. He used their travel-time curves to calculate in 1914 the depth of this boundary. These waves were named PcP for the P-waves reflected on the mantle-core boundary and ScS for the S-waves. In 1936, Inge Lehmann (1888-1993), a Danish seismologist, suggested the existence of a solid inner core based on the way P-waves are refracted on its surface. P-waves are absent between 105 and 142 degrees from an earthquake's epicenter because of the way they are refracted downward on the surface of the core (P-waves' shadow zone, Fig. 49). S-waves do not propagate through the liquid outer core and therefore S-waves do not reappear beyond 105 degrees (S-waves' shadow zone, Fig. 49). Many more categories of waves can be identified according to their travelling path inside the Earth (Fig. 50).

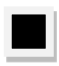

Figure 49: Paths of P-waves (left) and S-waves (right) inside the Earth after they originate from the focus of an earthquake. The distance from the earthquake's focus is expressed in degrees.

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* Notable exceptions to this trend are the Low Velocity Zone (LVZ) and the mantle-core boundary. See section 7.3 for more explanations.

Figure 50: Examples of seismic wave paths and their name (left) and examples of travel-time curves for different categories of waves (source: USGS).

4.3. The Earth's interior: the crust, the mantle, and the core

Based on the travel-time curves of seismic waves seen above, geologists constructed a model of the Earth's interior reflecting how the speed of P-waves and S-waves changes with depth (Fig. 51). The speed of seismic waves inferred from travel-time curves can be compared with the speed determined experimentally for different types of rocks. Using these data and other lines of evidence (e.g. field observations, meteorites…), geologists have established a model of the structure and composition of Earth's interior.

Figure 51: Variations of various properties of Earth's layers with increasing depth down to the center of the Earth. (A) Variations with depth of the temperature (geotherm, grey line) and melting temperature (pink line) of the mantle (down to 3000 km) and the core (from 3000 km to the center of the Earth), (B) variations with depth of the speed of seismic waves (P-wave in blue and S-wave in green) and density (black line), and (C) close-up of the variations with depth of S-wave speed from 0 to 900 km. Source: modified from Understanding Earth 6th edition.

4.3.1. The crust

The upper part of the continental crust is made primarily of felsic rocks (granite) whereas the oceanic crust is composed mainly of mafic rocks (basalt and gabbro). Felsic rocks are less dense than mafic rocks (see section 5.4). Differences in density and resistance to compression affect the speed of P-waves. As a result, seismic waves travel more slowly in felsic rocks than in mafic rocks. Seismic wave speed increases across the crust-mantle boundary (Moho) because the upper mantle is composed of peridotite (ultramafic rock) which is denser and more resistant to compression than the rocks of the continental and oceanic crusts.

The crust is thin under the ocean (7 km), thicker under continents (33 km) and thickest under large mountains (70 km). The variations of crust thickness relative to Earth's topography can be explained by the principle of ***isostasy***. Isostasy explains why the oceanic crust lies at low elevations below sea level (ocean basins) whereas the continental crust rises above sea level and, at places, reaches several kilometers in altitude (mountains).

A nice and easy way to explain this is to assimilate the mantle to a fluid and the crust to a solid floating on it. Although the mantle is solid in reality, it may "flow" (deform without breaking) on the very long term (over millions of years). In a static liquid, the pressure at a given depth is everywhere the same. The pressure is calculated by determining the weight of the material above the given depth (pressure = weight/area = $mg/A = \rho Vg/A = \rho gh$). As a result, the changing elevation of the crust floating on the mantle involves either changes in the crust thickness (Airy's model, Fig. 52A) or changes in the crust density (Pratt's model, Fig. 52B).

Figure 52: Two models explaining how the Earth's topography relates either to (A) lateral changes in the crust thickness (Airy's model) or (B) lateral changes in the crust density (Pratt's model). g = gravitational acceleration (in m/s^2), considered constant), ρ_m = density of the mantle (in kg/m^3), ρ_{cr} = density of the crust, ρ_w = density of seawater, w = ocean depth.

4.3.2. The mantle

The mantle down to 410 km is made of peridotite, a rock depleted in silica and enriched in iron and magnesium compared to crustal rocks (Fig. 51C). The most abundant mineral of the upper mantle is olivine. At a depth of around 410 km, the pressure forces olivine to adopt a more compact crystal structure. This change in mineralogy is responsible for a jump in wave speed at the same depth (Fig. 51C). A second —larger— jump occurs at around 660 km below the surface (Fig. 51C). This second jump corresponds to another change in the crystal structure of olivine with atoms becoming even more closely packed. The 660-km discontinuity defines the limit between the upper and lower mantle.

The outer rigid shell of the Earth is the lithosphere (100 km thick on average). The speed of seismic waves increase with increasing pressure (depth) in this layer but a remarkable thing happens near its base and in the upper part of the asthenosphere between 100 and 200 km below the surface: a decrease of wave speed defining a zone of low wave velocity (***Low Velocity Zone*** or LVZ, Fig. 51C). The slow down of seismic waves in this zone is due to the partial melting of the peridotite composing the upper mantle (see the geotherm crossing the mantle melting curve in Fig. 51A). This is the weak part of the asthenosphere on which the lithosphere (tectonic plates) may slide (see section 2 on plate tectonics).

4.3.3. The core

A complete change in composition and physical properties occurs across the mantle-core boundary. Above this boundary, the mantle is composed of solid silicate rocks. Below this boundary, the outer core is composed of a liquid iron and nickel alloy. This greatly affects the behavior of seismic waves. P-waves slow down abruptly whereas the speed of S-waves drops to zero (Fig. 51B). This mantle-core boundary also delineates a zone where temperature increases greatly. Scientists have suggested that part of the mantle directly above the mantle-core boundary could be melted and provide a source of magma that could be linked to hot spot volcanism.

Note that the core is probably composed mostly of iron and nickel[*] but its precise composition is still debated. The core could possibly also contain lighter elements like sulfur and oxygen.

[*] The composition of the core cannot be observed directly. Indirect evidence comes from the study of seismic waves, astronomical data (relative abundance of elements in the solar system, theory of planetary formation), and our knowledge of the composition of the mantle and the crust and the mass of the Earth. Another important piece of information comes from meteorites. One particular type of meteorites is composed almost entirely of iron and nickel. These meteorites are believed to be fragments of the core of other planetary bodies.

5. The climate system

5.1. Weather, climate, and components of the climate system

The weather is characterized by the atmospheric conditions (e.g. temperature, precipitations, cloud cover, wind speed) at a particular place at a particular time. It applies to short-term changes in these conditions (less than a day to a few weeks). Climate is the average atmospheric conditions of a region and applies to longer-term changes (years and longer). Climate can also apply to larger spatial scales like continents or oceans. One can even refer to the climate of a particular planet (i.e. average atmospheric conditions characterizing this planet). Factors influencing the climate include for example the latitude, the presence or absence of mountains, and the distance from the sea (continentality).

The climate system consists of five interacting natural components (Fig. 53, see also section 1) and is powered by ***solar energy***. These natural components are the atmosphere, the hydrosphere, the cryosphere, the lithosphere, and the biosphere. To these natural components, a sixth component can be added: the anthroposphere (sum of all factors related to human activities). The 2007 report of the Intergovernmental Panel on Climate Change (IPCC) stated: “there is *very high confidence* that the global average net effect of human activities since 1750 has been one of warming” and “most of the observed increase in global average temperature since the mid-20th century is *very likely* due to the observed increase in anthropogenic greenhouse gas concentrations”.

Figure 53: Schematic representation of Earth’s climate system and its components: the atmosphere, the hydrosphere, the cryosphere, the lithosphere, the biosphere, and the anthroposphere. Source: adapted from Earth’s climate: past and future, Ruddiman (2001).

5.2. The atmosphere

The atmosphere is the layer of gases surrounding the Earth. Its density decreases with increasing distance from Earth’s surface. The densest atmospheric layer is the troposphere and has an average thickness of 11 km (Fig. 54). That is where clouds form and atmospheric circulation takes place. It is a very mobile and fast-changing environment —think about how fast cloud cover may change and weather may sometimes shift from sunny to rainy in less than an hour—. Above the troposphere lies the stratosphere up to an altitude of 50 km. The stratosphere is a more stable environment. However, large volcanic eruptions can eject small particles in the stratosphere where they may remain several years before gravity removes them. This volcanic dust can potentially lower the amount of sunlight reaching the Earth’s surface. The stratosphere is also where the ozone (O_3) blocks harmful ultraviolet radiation from the Sun. Above the stratosphere, the atmosphere thins out progressively until interplanetary space is reached.

Figure 54: Vertical structure of the atmosphere. The grey line represents the evolution of air temperature with altitude. Source: adapted from Understanding Earth 6th edition.

Atmospheric circulation is driven by the uneven heating of Earth's surface (Fig. 55). The amount of incoming sunlight reaching the Earth's surface per unit area varies with latitude. At higher latitudes, sunlight reaches the Earth's surface at a higher angle and the same amount of solar energy must warm a larger surface than at lower latitudes (Fig. 55A). The amount of light reflected by Earth's surface is referred to as its ***albedo***. The albedo depends on the nature of the substrate and on the angle of incidence of the sunlight. Snow and ice have the largest albedos (60-90% of light reflected) whereas water has the lowest (5-10%). Also, the greater the angle of incidence, the greater the albedo (Fig. 55B). Therefore, a larger proportion of the incoming sunlight is absorbed by Earth's surface at lower latitudes, whereas a larger proportion is reflected at higher latitudes. Cloud cover should also be taken into account as some of the incident sunlight is reflected by the clouds. All in all, the poles reflect more sunlight than the other regions of the globe (Fig. 55C). A maximum solar energy is absorbed at the equator and a minimum at the poles. Therefore, the poles are much colder than the equator.

Figure 55: Factors contributing to the unequal heating of the Earth's surface. (A) the same amount of solar radiation heats a larger area of Earth's surface at higher latitudes, (B) the relationship between the latitude, the angle of incidence of the sunlight, and the proportion of light absorbed/reflected, and (C) graph showing the unequal heating of Earth's surface and the heat loss at the poles vs. the maximum heat gain at the equator. Source: adapted from Earth's climate: past and future, Ruddiman (2001).

As a result, Polar Regions lose heat whereas low-latitudes areas gain heat. Since on average the poles do not get cooler and the equator does not get warmer over time, there must be a transport of heat from the equator to the poles. Heat is transported from low to high latitudes in the air (atmospheric circulation) and in the oceans (oceanic circulation).

The circulation of air and water masses does not follow a straight line from the equator to the poles. Their trajectory is complicated by the Earth's rotation and constrained by the topography of ocean basins and landmasses. To understand the effect of Earth's rotation, let's first consider the example of a fixed object (Fig. 56A). If this object is situated at one of the poles, it will spin around on itself (with no eastward velocity). As the object's position gets farther away from the poles, it will move in the same direction of Earth's rotation faster and faster (its eastward velocity increases). Now let's consider the effect of Earth's rotation on a moving object, say a projectile launched from France in the direction of the equator (Fig. 56B). In the case of a non-rotating Earth, the object would follow a longitudinal line (Fig. 56B-1). In reality, the projectile seen from the perspective of an Earth-bound observer is not following a straight line but is deflected to the right (Fig. 56B-2). This is because Paris is moving eastward more slowly than any point situated on the equator. If the projectile had been launched from somewhere in the southern hemisphere toward the equator, its trajectory would have been deflected to the left. This is due to the ***Coriolis effect***. The Coriolis effect influences the trajectory of any objects or fluids moving in any direction in the air, in the ocean, and even in the liquid outer core. For air or water masses in motion, the result is a deviation to the right in the northern hemisphere and to the left in the southern hemisphere. The Coriolis effect becomes significant for any object traveling over long distances but it affects moving objects at all scales. For instance, “a ball thrown horizontally 100 m in 4 s in the United States will, due to the Coriolis force, deviate 1.5 cm to the right” (quoted from Persson, 1998).

Figure 56: Illustration of the Coriolis effect. The eastward velocity of a fixed point on Earth is maximum at the equator and 0 at the poles (A). From the viewpoint of an Earth-bound observer, an object launched from Paris toward the equator would follow a longitudinal line if the Earth was not rotating (B-1) and follows a curved path deflected to the right in the case of a rotating Earth (B-2). The deflection occurs because the object leaves Paris with an eastward velocity lower than the eastward velocity of the targeted point on the equator. The deflection would be to the left if the object was launched from a location in the southern hemisphere.

Let’s begin our description of the atmospheric circulation by the equatorial region (Fig. 57). In this area, solar heating of Earth’s surface is maximum. Hot, low-density air rises high in the troposphere, producing a low atmospheric pressure belt. As the rising air cools, water vapor — coming from the evaporation of ocean water— condenses and large clouds form, resulting in abundant rainfalls. These rainfalls sustain highly productive rainforests in the equatorial region (e.g. Amazon, Ituri). As the air moves away from the equator at high altitude, it cools further, becomes denser, and sinks around 30º of latitude (area of high atmospheric pressure). Closer to the ground, air rushes toward the equator to replace the hot rising air. This produces the ***trade winds*** which blow west due to the Coriolis force and transport abundant water vapor that feeds the large equatorial clouds. The area where northern and southern trade winds meet is called the ***intertropical convergence zone*** (ITCZ). The combination of the high-altitude movement of air away from the equator and the surface trade winds blowing toward the equator represents a component of the global atmospheric circulation called the ***Hadley Cell*** (Fig. 57).

Near 30º of latitude, dry air sinks and becomes warmer due to the increase in atmospheric pressure at lower altitudes (think about the increase in temperature when you compress air in a bicycle pump). The sinking air is dry because most of its water vapor was released as rainfalls near the equator. Rainfalls are therefore limited in this region. What makes rainfalls even less likely is the increase in temperature which allows the air to hold more water vapor. Consequently these high-pressure subtropical areas are characterized by an arid climate (e.g. Sahara, Kalahari, Gobi, Australian desert). Part of the sinking air returns to the equatorial region (carried by the trade winds) and another part moves poleward. This poleward flow is deflected by the Coriolis force and produces winds called the ***westerlies*** which transport heat from low to high latitudes (blowing in a

SW-NE direction in the northern hemisphere).

The ***polar front*** is a zone where the warm air carried by westerlies meets the much colder air moving away from the poles. The warm air rises above the cold, denser air and produces clouds and rainfalls as water vapor condenses at high altitudes. North of the polar front, the cold polar winds are called ***polar easterlies*** (blowing from the NE in the northern hemisphere).

Figure 57: Simplified representation of Earth's atmospheric circulation. Note that other models include two other cells, one at mid-latitudes and another at high latitudes. Source: adapted from Earth's climate: past and future, Ruddiman (2001)

5.3. The hydrosphere

The oceans too carry heat from the tropics to the poles. There are two types of ocean circulation involved in this heat transport: the ***surface*** —wind-driven— ***circulation*** (Fig. 58) and the ***thermohaline circulation*** (Fig. 59). The former results from the action of winds that push water masses horizontally and affects the top 100 m of the water column. Because of the Coriolis force, surface currents are deflected to the right of the wind direction in the northern hemisphere and to the left in the southern hemisphere. As a result of this and of the topography of ocean basins and landmasses, surface currents delineate large ***gyres*** —clockwise in the northern hemisphere and anticlockwise in the southern hemisphere—. Ocean currents are not limited to the surface. There is also a slow oceanic circulation driven by differences in water density which affects shallow and deep water masses. Ocean water density is a function of temperature and salinity, hence the name thermohaline circulation (*therme* = heat, *halos* = salt). There are areas at high latitudes where dense ocean water (salty and cold) sinks toward the bottom of the ocean (***downwelling***). A major source of deep water is the North Atlantic region near Greenland where water becomes saltier due to evaporation and is chilled by cold winds. The cold, dense water sinks to the bottom of the Atlantic and starts its deep journey in the direction of the southern hemisphere. Another area of deep water formation is the South Atlantic region near Antarctica (e.g. the Weddell Sea, not shown in Fig. 59). There is no deep water formation in the Pacific. Since water masses sink in some areas, there must be other areas where water masses return to the surface (***upwelling***). Upwelling can occur along the equator where trade winds push surface waters in opposite directions (Fig. 60B). It can also occur along the west coast of America and Africa where surface winds blowing along the coast push water masses offshore (Fig. 60A). The thermohaline circulation is very slow. It takes on average more than one thousand years for ocean water to return to the surface after it sinks in the North Atlantic.

Figure 58: Surface, wind-driven circulation. Red lines: warm surface currents, blue lines: cold surface currents. Source: Understanding Earth 6th edition.

Figure 59: Thermohaline circulation. Note that this map is highly simplified and that all the downwelling and upwelling areas are not shown. Source: Understanding Earth 6th edition

Figure 60: Upwelling areas. (A) Upwelling along a coast where winds push surface water away from the coast and (B) upwelling along the equator where winds push surface water away from the equator. Source: Earth's climate: past and future, Ruddiman (2001).

5.4. The cryosphere

The cryosphere comprises the Antarctic and Arctic ice sheets (Fig. 61), continental glaciers and snow, sea ice, and permafrost (permanently frozen soils of high-latitude regions). The cryosphere influences the Earth's climate because of the high albedo of snow and ice and because of its influence on the thermohaline circulation.

Figure 61: Extent of the Antarctic ice sheet (left) and arctic ice sheet (right) in summer. Source: NASA.

To understand its influence on the thermohaline circulation, think about the impact of the melting of the Greenland ice sheet in response to global warming. More fresh water added to the North Atlantic ocean would decrease the salinity —hence the density— of surface water and prevent it from sinking. This would slow down or even shut down the thermohaline circulation. The consequence would be an alteration of the heat transfer between low and high latitudes in the North Atlantic. It would have a great impact on the climate, perhaps leading to a significant cooling of Western Europe. Questions related to the current state of the thermohaline circulation and the impact a slowdown (or shutdown) of the thermohaline circulation would have on the climate are particularly relevant in the present context of global warming.

5.5. The lithosphere

The lithosphere influences the climate in a number of ways at various spatial and time scales. The nature of the land surface determines the albedo. In addition, the land tends to warm and cool more quickly than the ocean. Large bodies of water, particularly the oceans, can store a large amount of heat in summer and release it slowly in winter. Oceans tend to buffer changes in surface temperature. As a consequence, the annual range of surface temperature of inland continental areas is much larger than along the coast or offshore.

In the tropics, the difference between the thermal properties of the land and the ocean gives rise to the ***monsoonal cycle*** (e.g. Indian monsoon). In summer, the land warms quickly whereas the ocean remains relatively cold. Warm air above the land rises and draws humid air masses in from the ocean. The air rises, condenses and produces abundant rainfalls (Fig. 62A). In winter, the land cools more quickly than the ocean. The air tends to sink over the land. It then flows toward the ocean where it warms, becomes humid, rises and produces rainfalls offshore (Fig. 62B).

Figure 62: Monsoonal cycle. (A) Summer monsoon and (B) winter monsoon. See text for explanations. Source: Earth's climate: past and future, Ruddiman (2001).

The effect of land topography is also crucial. When warm air that has swept over the ocean and taken up abundant water vapor meets a mountain chain, the air rises and water vapor condenses, forming clouds and producing rainfalls (orographic precipitation). The air that flows down the opposite flank is dry. This results in a low-precipitation area on the leeward side (opposite to the windward side), a phenomenon referred to as the ***rain shadow effect*** (Fig. 63).

Figure 63: Orographic precipitation and rain shadow effect. See text for explanations. Source: Earth's climate: past and future, Ruddiman (2001).

Volcanoes release greenhouse gases in the atmosphere that can influence the climate (e.g. CO_2 and water vapor, see section 8.8. for more information on the greenhouse effect). Large volcanic eruptions can also produce a tremendous amount of dusts which may block part of the incoming sunlight.
On a much larger time scale (over millions of years), plate tectonics controls the topography and shape of ocean basins and landmasses. Plate tectonics therefore controls ocean circulation patterns and influences the transfer of heat from low to high latitudes. The absence of a landmass at the poles makes it more difficult for snow and ice to accumulate and build up ice sheets.

5.6. The biosphere

Living organisms have a crucial influence on the uptake and release of greenhouse gases. Using photosynthesis to fix inorganic carbon (CO_2), terrestrial and marine plants store a large amount of carbon derived from carbon dioxide. Some of this carbon dioxide is returned to the atmosphere during cellular respiration.

Photosynthesis: $6H_2O + 6CO_2 + \text{solar energy} \rightarrow C_6H_{12}O_6 + 6O_2$

Cellular respiration: $C_6H_{12}O_6 + 6O_2 \rightarrow 6H_2O + 6CO_2 + \text{chemical energy}$

Transpiration, the evaporation of excess water excreted by plants, accounts for as much as 10% of the moisture in the atmosphere, the other 90% being linked to evaporation of the oceans (source:

USGS). ***Evapotranspiration*** (the sum of evaporation and plant transpiration) influences air temperature and the amount of precipitation. The transformation of liquid water to water vapor requires energy and this energy is taken from the environment as heat. Heat is absorbed by evaporating water which tends to decrease the temperature of the environment, hence the cooling effect of plant transpiration (and sweating!). In addition, the water evaporating from plants can condense at high altitude and produce rain, influencing local precipitation patterns.

5.7. Earth's radiation balance

The source of energy driving the climate system is the sunlight (external forcing). The energy radiated by the Sun and reaching the Earth consists mostly of visible light (***shortwave radiation***). The average amount of solar radiation arriving on top of the atmosphere is 342 W/m2 (1 W = 1 Joule per second). What happens to this incoming radiation? (Fig. 64)

70% (240 W/m2) is absorbed by Earth (30% by the clouds and 70% by Earth's surface) and 30% is reflected/scattered back into space (85% by the top of clouds and 15% by Earth's surface). The 240 W/m2 absorbed by the Earth drives the climate system.

Because Earth's temperature is constant, the energy absorbed (240 W/m2) must be radiated back into space. In other words, the radiation balance must be maintained. Any object warmer than absolute zero (-273ºC) emits light. In the case of Earth, the light that is radiated by Earth's surface and by the top of clouds is in the infrared range (***longwave radiation***). What happens to the infrared light emitted by Earth's surface?

The amount of radiation sent back into space (240 W/m2) comes mostly from the infrared light emitted by the top of clouds. Very little of the radiation emitted by Earth's surface makes it into space (only 5%). If we translate the infrared light radiated back into space (240 W/m2) in terms of temperature, we would obtain an average surface temperature of -16ºC. That is far below the actual value of +15ºC at Earth's surface! The difference is due to the infrared light that is absorbed and re-emitted back to Earth's surface by clouds and greenhouse gases. This makes the Earth's surface 31ºC warmer than it would be in the absence of greenhouse effect! Note the dual effect of clouds on climate: (1) cooling due to the reflection of the incoming solar radiation and (2) warming because clouds re-emit infrared light back to Earth's surface. The most important greenhouse gases are water vapor, carbon dioxide (CO_2) and methane (CH_4). In the following two chapters of this course, we will talk more about the impact of human activities on the concentration of greenhouse gases in the atmosphere and their influence on climate.

So far in our discussion on the Earth's heat budget, we have been talking about heat transfer involving thermal radiation (solar radiation in the visible range of light and radiation emitted by Earth's surface and the clouds in the infrared range). There are two other important processes contributing to heat transfer that influence the climate system:

(1) One involves a transport of heat by moving air masses. Hot, low-density air masses tend to move upward whereas cold, high-density air masses tend to move downward. This vertical movement of air masses driven by density differences is called ***convection*** (same process that drives movements of matter in the asthenospheric mantle). Heat carried by moving air masses is called ***sensible heat***.

(2) The other process involves heat released during condensation of water vapor and heat absorbed during evaporation of liquid water. Condensation of water vapor leads to a warming of the environment, whereas evaporation of liquid water leads to a cooling of the environment (remember the effect of plants transpiration). This heat is called ***latent heat***.

Figure 64: Earth's radiation budget. See text for explanation. Source: Earth's climate: past and future, Ruddiman (2001).

6. The water cycle

See Slides

7. The carbon cycle

7.1. Box model of the carbon cycle

Without the greenhouse effect, our planet would experience a permanent ice age and life as we know it would not be possible. The main contributors to the greenhouse effect are the clouds and greenhouse gases. The three main greenhouse gases are water vapor, carbon dioxide (CO_2), and methane (CH_4). Human activities, particularly the burning of fossil fuels (oil, coal, natural gases), have led to a significant increase of CO_2 in the atmosphere which is warming the planet faster than would be expected without the addition of this anthropogenic CO_2. To understand the climate and predict its evolution in the future, it is essential to understand the natural mechanisms controlling the concentrations of CO_2 and CH_4 in the atmosphere. We should know where the carbon comes from and how it is removed from the atmosphere. We should also be able to quantify how much carbon flows in (***input flux***) and how much flows out (***output flux***), and how much time on average it stays in the atmosphere (***residence time***). The atmosphere is one ***reservoir*** of carbon but we need to identify the other reservoirs and describe how carbon moves from one to another. An obvious reservoir of carbon is the biosphere. Carbon atoms are essential for life because they form the backbones of large organic molecules. Other important reservoirs of carbon are the oceans, carbonate rocks (mainly those composed of calcium carbonate, i.e. $CaCO_3$), and fossil fuels (coal, oil, natural gas). In any of these reservoirs, the amount of carbon (***reservoir size***) has fluctuated over geological times. When there is a natural balance between what enters and leaves each reservoir, the system is said to be in a ***steady state***. It means that for any given reservoir the input flux is equal to the output flux (often expressed in gigaton per year or Gt/yr) and the reservoir size remains constant. In this condition, the residence time is the reservoir size divided by the input flux (or the output flux). Hence, the carbon cycle can be represented in a simplified manner by various reservoirs which exchange certain amounts of carbon at certain rates. This representation of the carbon cycle is called a ***box model***. The carbon cycle is one example of biogeochemical cycles that can be represented using a box model. Other examples are the water cycle, the nitrogen cycle, the sulfur cycle... Box models are particularly useful for modeling biogeochemical cycles using computer programs.

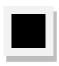


Figure 65: Reservoir in a steady state condition where the input flux is equal to the output flux and the reservoir size remains constant.

7.2. The natural carbon cycle

A simplified box model of the natural carbon cycle (without the influence of human activities) is shown in figure 66.

Figure 66: Simplified box model of the natural carbon cycle. The size of each reservoir is expressed in gigaton (Gt). Input and output fluxes are expressed in gigaton/year (Gt/yr). DOC means dissolved organic carbon and represents organic molecules present marine (oceans) and freshwater (rivers and lakes) systems. Source: partly derived from Biogeochemistry: An Analysis of Global Change (2013).

If we examine the figure above and examine the fluxes of carbon from one reservoir to another, several interesting observations can be made:

- Most carbon is stored in the reservoir "carbonate rocks".
- The largest fluxes of carbon are between the land surface (plants and soils) and the atmosphere and between the atmosphere and the ocean ("fast domain").
- Soils store three times as much carbon as plants.
- The smallest fluxes are related to calcium carbonate deposition in the ocean, calcium carbonate dissolution and weathering of silicate rocks, volcanism, and carbon fluxes in and out of the fossil fuel reservoir ("slow domain").
- The residence time of carbon in the atmosphere is very short (3.5 years) whereas in carbonate rocks the residence time is very long (6×10^8 years).

7.2.1. Land surface (plants, soils) ↔ atmosphere

The greatest flux of carbon is the exchange of carbon between the terrestrial biosphere and the atmosphere. This involves primarily ***plant photosynthesis*** (I), ***cellular respiration*** (II), and the ***decomposition of organic matter*** by microbes (III). During photosynthesis, plants take CO_2 from the atmosphere to build organic molecules using sunlight as a source of energy. Part of this CO_2 returns in the atmosphere during cellular respiration, a reaction producing energy. When plants die, organic matter is incorporated in soils where it is decomposed by microbes. This process releases CO_2 in the atmosphere and is one component of soil respiration. Another important component of soil respiration is cellular respiration of plant roots which also releases CO_2 in the atmosphere.

Photosynthesis

$$6H_2O + 6CO_2 + \text{sunlight} \rightarrow C_6H_{12}O_6 + 6O_2 \qquad \text{(I)}$$

Cellular respiration

$$C_6H_{12}O_6 + 6O_2 \rightarrow 6H_2O + 6CO_2 + \text{energy} \qquad \text{(II)}$$

Decomposition of organic matter by microbes

a. *Aerobic*

$$C_6H_{12}O_6 + 6O_2 \rightarrow 6H_2O + 6CO_2 + \text{energy}$$

(III)

b. *Anaerobic*

$$\text{Organic matter} \xrightarrow{\text{Fermentation}} \text{Acetate} \xrightarrow{\text{Methanogenesis}} CO_2 + CH_4$$

$$CO_2 + 4H_2 \rightarrow CH_4 + 2H_2O$$

7.2.2. Atmosphere ↔ ocean

Understanding how the oceans absorb the atmospheric CO_2 is essential to predict the future of Earth's climate. The local flux of CO_2 between the atmosphere and the ocean depends on various factors including seawater composition, seawater and air temperature, and winds. Gases are less soluble at higher temperature (a soda drink becomes flat faster at room temperature than in the fridge). Increasing temperature reduces CO_2 solubility in the ocean. Stronger winds increase the exchange of CO_2 between the atmosphere and the ocean because of the formation of air bubbles in breaking waves and the production of sea spray (droplets of seawater blown from waves). Bubbles and sea spray increase the contact surface between the air and seawater and therefore enhance air-sea gas exchange.

7.2.3. Oceans ↔ marine life ↔ carbonate rocks

Marine photosynthetic organisms use the dissolved CO_2 in the ocean as a source of carbon they incorporate in their tissue. Marine organisms also perform cellular respiration which release CO_2 in the environment. Some of the organic carbon is recycled in the surface ocean and some is transferred to the deep ocean ("marine snow"). Much like on land organic matter is decomposed by microbes which releases CO_2 in the environment. In addition, many marine organisms are able to precipitate $CaCO_3$ (calcium carbonate) to build a hard shell or skeleton (Fig. 67). Hard parts of marine organisms made of $CaCO_3$ accumulate on the sea floor and enter the carbonate reservoir (but see slides for a definition of lysocline and carbonate compensation depth). The process of calcification releases CO_2 in the environment as shown by the following reaction:

$$Ca^{2+} + 2HCO_3^- \rightarrow CaCO_3 + H_2O + CO_2$$

The rate at which carbon enters the carbonate reservoir is very low (0.2Gt/yr). However, the carbon stored in this reservoir may remain there for millions of years. In addition, the organic matter that escaped decomposition can be buried and stored in the lithosphere for a very long time.

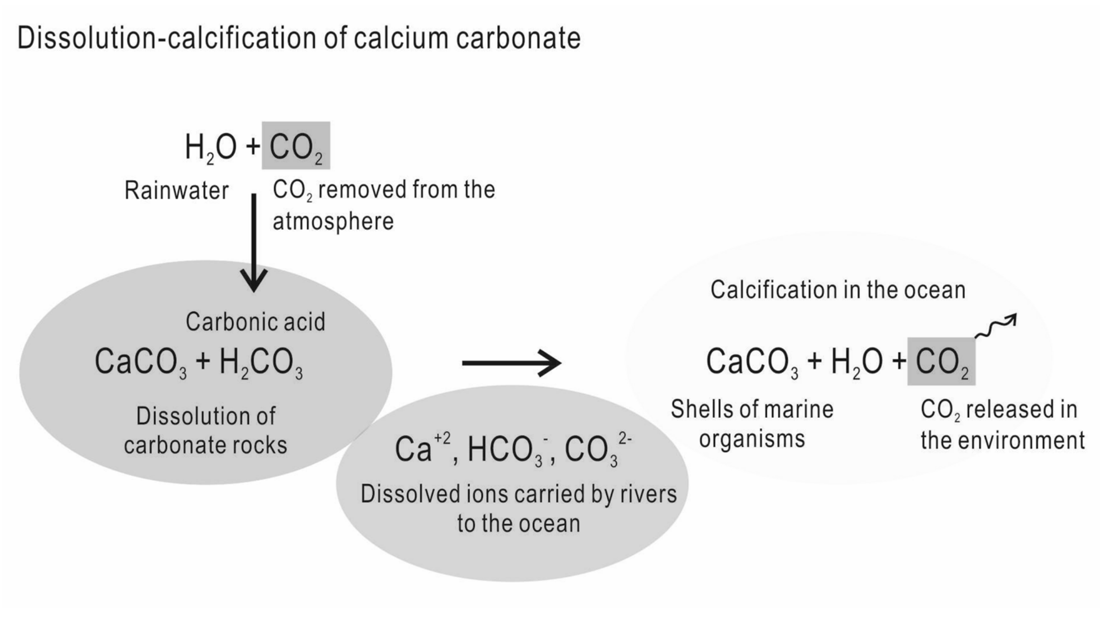
Figure 67: Examples of marine organisms producing a shell or skeleton made of calcium carbonate. (A) Live benthic foraminifera *Calcarina* (source: Catalogue of organisms, http://coo.fieldofscience.com), (B) sediment composed of *Calcarina* shells (source: Wikipedia), (C) live calcareous green algae *Halimeda* (source: www.algaebase.org), (D) sediment composed of *Halimeda* segments, (E) coral reef built primarily by corals and coralline algae (Great Barrier Reef of Australia, source: AIMS), (F) encrusting coralline algae (source: Encyclopedia of Earth, www.eoearth.org), and (G) detail of a coral skeleton.

7.2.4. Lithosphere ↔ atmosphere / ocean

The lithosphere is the largest reservoir of carbon. The carbon present in the lithosphere occurs primarily in carbonate rocks, organic matter, and carbon dioxide. Some important processes behind the exchange of carbon between the lithosphere and the atmosphere (or the ocean) are explained below.

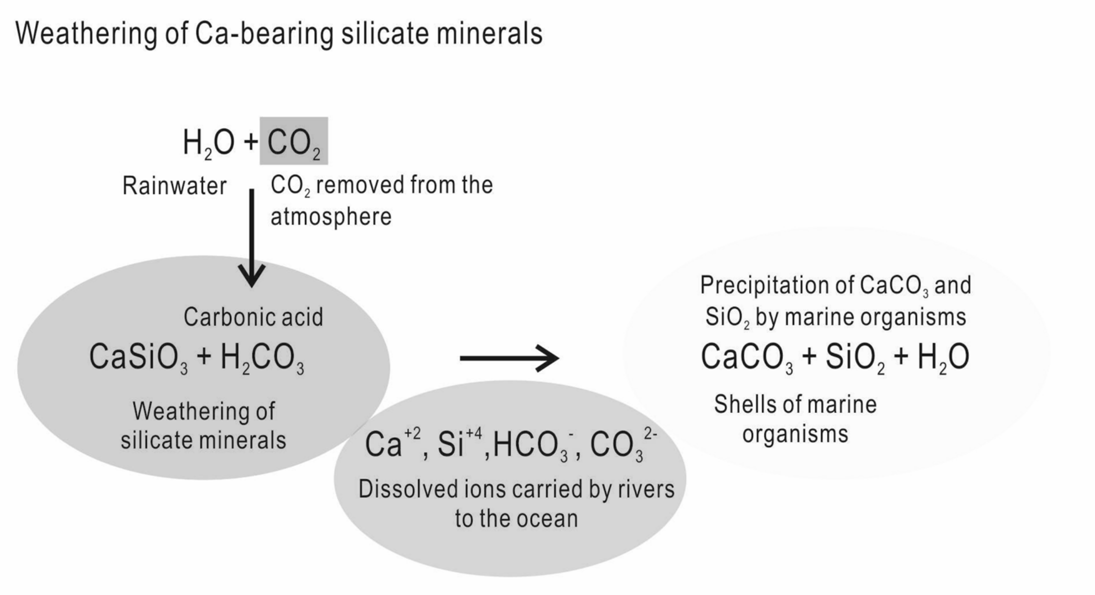
a. ***Dissolution / precipitation of calcium carbonate***

Exchange of carbon between the atmosphere and the lithosphere occurs during dissolution and precipitation of $CaCO_3$. 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, calcifying marine organisms use these ions to build their hard parts (shells, skeletons…). The process of dissolution of calcium carbonate removes CO_2 from the atmosphere. However, the same amount of CO_2 is released during calcification. Therefore, the overall process of dissolution-calcification does not result in a net loss (nor gain) of CO_2 in the atmosphere as shown by the following simplified reactions (from Ruddiman, 2001):



b. ***Weathering of silicate rocks***

An important process controlling the long-term concentration of atmospheric CO_2 (over millions of years) is the weathering of silicate rocks. Silicate minerals are the most abundant minerals in the Earth's crust (see section 2, page 3). Here we are concerned more specifically with Ca-bearing silicate minerals. Like carbonate rocks, Ca-bearing silicate minerals 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-}, H^+) and clays. These ions are transported by rivers to the ocean and some can be used by marine organisms to build hard parts made of $CaCO_3$ and SiO_2*. We can use the simple formula $CaSiO_3$ to represent Ca-bearing silicate minerals and the overall process of silicate weathering combined with calcification and silification in the ocean can be summarized by the following reactions (from Ruddiman, 2001):



Most importantly, the above reactions show that the weathering of silicate rocks results in a net removal of CO_2 from the atmosphere! This slow removal of CO_2 must be balanced by an equally slow input of CO_2. This input of CO_2 is provided by volcanoes and hot springs.

c. ***Volcanism***

The CO_2 from Earth's interior is released by volcanoes and hot springs. The rate of CO_2 input by volcanic activity is variable and depends on the frequency and intensity of volcanic eruptions. The geological record shows evidence of past volcanic eruptions which were far more intense and lasted far longer than those recorded in human history. Such events can release a huge amount of CO_2 in the atmosphere and potentially alter the Earth's climate for an extensive period of time. An example of exceptionally intense period of volcanism is documented in the Siberian Traps, a succession of huge basaltic lava flows covering millions of km^2 and related to an eruptive event spanning one million year that happened 250-251 million years ago. This event may have been one of the factors which triggered the largest mass extinction in life history, i.e. the end-Permian mass extinction.

* Silica, i.e. SiO_2, is with $CaCO_3$ a major component of mineral hard parts produced by marine organisms. Examples of organisms producing hard parts made of SiO_2 are diatoms (phytoplankton), radiolarians (zooplankton) and siliceous sponges.

d. ***Fossil fuels***

Another reservoir of carbon which is part of the lithosphere is the organic matter preserved in rock strata. A common sedimentary rock composed of the accumulation of organic matter is coal. Other products of the accumulation of organic matter are oil and natural gas (mostly methane). In the natural carbon cycle (without the influence of human activities), the transfer of carbon from the reservoir of fossil organic carbon to the atmosphere is extremely slow. The organic carbon present in the lithosphere can return into the atmosphere via volcanic activity, natural fire and oxidation (biological or not).

In conclusion, unlike the exchange of carbon between the biosphere and the atmosphere, the natural rate at which carbon is exchanged between the lithosphere and the atmosphere is extremely slow. The two main processes controlling the flux of carbon between the lithosphere and the atmosphere are silicate weathering and volcanism. These very slow exchanges affect the concentration of atmospheric CO_2 (hence climate!) over the very long term (millions of years) and are controlled by plate tectonics (see slides and text of chapter 10).

7.3. Anthropogenic perturbations, global warming, and ocean acidification

In May 2013, the level of atmospheric CO_2 has reached 400 ppm (parts per million) for the first time since accurate measurements began in 1958 (Fig. 68). The rapid increase in atmospheric CO_2 is related to human activities, primarily the combustion of fossil fuels. Figure 69 shows the impact of human activities on the carbon cycle. Besides fossil fuel burning, other human-induced sources of atmospheric CO_2 are related to deforestation and cement production. The figure also indicates that natural processes removing CO_2 from the atmosphere (plant growth and ocean uptake) do not counterbalanced human emissions, leading to a net increase in atmospheric CO_2 (3.2 ± 0.1 Gt/yr for the 1990s and 4.1 ± 0.1 Gt/yr for 2000-2005, source: IPCC 2007 report).
We know with a high level of certainty that human emissions of CO_2 contributes in part to the present-day global warming. What we don’t know for sure is exactly how much warming is caused by anthropogenic CO_2. As we will see in the next chapter, the Earth has experienced many cold/glacial and warm/interglacial periods over the last 2.5 million years. The last glacial period reached its coldest peak 20,000 years ago when the extent of ice sheets was maximum. Since this last ice age, the Earth is warming again and ice sheets are retreating. Warming caused by humans is accelerating this natural trend.

Figure 68: Evolution of the concentration of atmospheric CO2 for the past 55 years. Source: Scripps Institution of Oceanography, CO_2 Program, Mauna Loa Observatory.

Figure 69: Impact of human activities on the carbon cycle. Carbon fluxes related to human activities are partly derived from Biogeochemistry: An Analysis of Global Change (2013).

We will discuss more about the impact of human activities on climate later in this course. Now let's briefly examine another consequence of the increase of atmospheric CO_2: **ocean acidification**.

In the ocean, CO_2 reacts with H_2O to form carbonic acid (H_2CO_3). Most of the carbonic acid dissociates to hydrogen ions (H^+) and bicarbonate ions (HCO_3^-). Some of the bicarbonate ions dissociate to hydrogen ions and carbonate ions (CO_3^{2-}). Adding CO_2 in seawater increases the concentration of hydrogen ions and therefore decreases the pH. The chemical reactions involved can be expressed as follows:

$$CO_2 + H_2O \leftrightarrow H_2CO_3$$

$$H_2CO_3 \leftrightarrow H^+ + HCO_3^-$$

$$HCO_3^- \leftrightarrow H^+ + CO_3^{2-}$$

Since these reactions involves the concentration of bicarbonate and carbonate ions in the ocean, it is clear that increasing the amount of CO_2 absorbed by the ocean has an impact on biological calcification (e.g. corals and coral reef ecosystems). One way to look at the problem is to consider that organisms combine calcium ions (Ca^{2+}) and carbonate ions (CO_3^{2-}) to produce their shell or skeleton made of calcium carbonate ($CaCO_3$). What would happen to the concentration of carbonate ions if more CO_2 was added in the ocean? As more CO_2 is added, more hydrogen ions is produced and pH would decrease. Some of these hydrogen ions would tend to bond with carbonate ions to form bicarbonate. Therefore, it would also lead to a decrease in the concentration of carbonate ions and would potentially make calcification more difficult (since carbonate ions are needed for calcification!). The relationship between pH and the concentration of dissolved CO_2, HCO_3^- and CO_3^{2-} is shown in figure 70. It clearly shows the decrease in concentration of CO_3^{2-} with decreasing pH. Several experiments have shown that adding CO_2 to the ocean would be detrimental to calcifying organisms. However, biological calcification is a complex process which is far from being well understood. More work is needed to predict the impact of ocean acidification on calcifying organisms and related ecosystems.

Figure 70: Concentrations of the dissolved CO_2, bicarbonate and carbonate ions as a function of pH. Source: Ridgwell and Zeebe (2005).

8. Climate changes

8.1. Short-term regional variations

By short-term climate changes, we refer here to changes occurring over years to decades. Over this timescale, climate is influenced by interactions between the atmosphere, the sea, and the land surface. In chapter 8, we already mentioned the monsoonal cycle which is driven by land-sea interactions. Another example of short-term climate variation is the phenomenon called ***El Niño***. El Nino refers to a cyclic warming of the eastern equatorial Pacific Ocean that has a variable periodicity of 3 to 7 years and lasts approximately 1 year. A cooling of the same region sometimes follows El Niño events and is called ***La Niña***. In “normal” conditions, the trade winds push water westward away from the coast of western South America (Fig. 71). This movement of water away from the coast creates a strong coastal upwelling that brings deep, cold, nutrient-rich water to the surface, replacing the water mass moving off shore. This situation is very beneficial for the fishing industry along the coast of western South America (e.g. Peru, Ecuador). Surface waters are much warmer in the western Pacific where hot air masses rise, water vapor condenses into clouds associated with heavy rainfalls (low pressure system, e.g. rainforest of Indonesia). In the eastern Pacific, surface waters are cooler and rainfalls are less frequent (high pressure system, e.g. Sechura desert in Peru).

Figure 71: Map of the equatorial Pacific Ocean showing the distribution of sea surface temperatures in “normal” atmospheric and oceanographic conditions. Below the map is a cross section of the same area showing the temperature profile of the water column, the upwelling of deep water (blue arrows) along the coast of western South America, and rainfalls in the western Pacific Ocean (by opposition to the colder, drier climate of the eastern Pacific). Source: adapted from NOAA.

During El Niño events, trade winds weaken or even start blowing in the opposite direction. As a consequence, the upwelling of cold water along the coast of western South America ceases and the eastern Pacific warms (Fig. 72B). This situation results in a collapse of fish populations. El Niño events are therefore detrimental to the eastern Pacific fisheries. The warming of the eastern Pacific also increases the frequency of rainfalls and thunderstorms, with possible flash floods occurring in countries bordering the eastern Pacific. The counterpart of El Niño, La Niña, is characterized by a strengthening of the trade winds, which enhances the upwelling of cold, nutrient-rich water along the coast of western South America (Fig. 72C). This results in a cooler eastern Pacific and warmer western Pacific (the opposite of El Niño). La Niña has a positive impact on the fishing industry along the coast of western South America. Variations in atmospheric and oceanographic conditions generating El Niño and La Niñia events are referred to as the ***El Niño-Southern Oscillation*** (***ENSO***).

Figure 72: Map of the equatorial Pacific Ocean showing the distribution of sea surface temperatures during (A) “normal” conditions, (B) El Niño, and (C) La Niña.

8.2. Long-term global variations

8.2.1. Pleistocene: The last 2,000,000 years

The climate of the past 2 million years has been characterized by an alternation of warm and cold periods (Fig. 73). Cold ***glacial periods*** (ice ages) are characterized by a lower global average temperature and more extensive polar ice sheets. More ice and snow stored on continents means less water in the oceans and a lower average sea level. Warm ***interglacial periods*** are characterized by a higher global average temperature, less extensive ice sheets, and a higher average sea level. The highs and lows of figure 73 delineate an asymmetric sawtooth pattern characterized by successive periods of rapid warming and slow cooling.

Figure 73: Curve showing the succession of glacial (lows) and interglacial (highs) periods over the past 1.8 million years. Values plotted along the vertical axis ($\delta^{18}O$) are derived from the ratio $^{18}O/^{16}O$ measured in shells of marine microorganisms in sediment cores taken at various locations (primarily in the Atlantic Ocean). The relationship between $^{18}O/^{16}O$ and the volume of ice sheets is explained in section 10.3 of this chapter. Source: $\delta^{18}O$ curve from Lisiecki and Raymo (2005).

What controls glacial-interglacial cycles?

Periodic variations in the amount of solar radiation reaching the Earth are thought to be a major controlling factor of glacial-interglacial cycles. These long-term variations in Earth’s insolation are caused by periodic changes in Earth’s movement around the Sun. The first person who identified these cycles is a Serbian scientist named Milutin Milankovitch (1879-1958). Three types of orbital changes are involved. They are known as the Milankovitch cycles:

- ***Eccentricity*** cycle (Fig. 74A): Earth’s orbit around the Sun changes from an elliptical shape to a nearly circular shape with a period of 100,000 years (superimposed to a longer cycle of 400,000 years).
- ***Obliquity*** (or tilt) cycle (Fig. 74B): The angle between Earth’s rotation axis and its orbital plane (obliquity or tilt) varies between 22.1 and 24.5 degrees with a period of 41,000 years
- ***Precession*** cycle (Fig. 74C): The wobble of Earth’s rotation axis is called precession and has a period of 23,000 years.

Figure 73 shows that the frequency and amplitude of climatic fluctuations have not always been the same. Comparing the climate record before and after 800,000 years in figure 73 shows that the frequency of glacial and interglacial periods has decreased and their amplitude has increased over time. A closer examination of figure 73 shows that the periodicity of glacial-interglacial cycles shifted from 41,000 years to 100,000 years, suggesting that they were first driven mainly by Earth's obliquity cycle and then became more influenced by Earth's eccentricity cycle.

Figure 74: The Milankovitch cycles. Source: adapted from Zachos et al. (2001).

Figure 75: Variations over the past 400,000 years of various parameters derived from the analysis of air bubbles trapped in ice cores from the Antarctic ice sheet (curves **a** to **d**) and of insolation (derived from Milankovitch cycles, curve **e**). **a**: concentration of CO_2, **b**: temperature (partly derived from the ratio $^{18}O/^{16}O$), **c**: concentration of CH_4 (methane), and **d**: $\delta^{18}O$ (parameter derived from the ratio $^{18}O/^{16}O$ of O_2*). Source: Petit et al. (1999).

The Milankovitch cycles alone cannot explain the amplitude of climate changes nor how abruptly glacial periods end. There must be positive feedback mechanisms enhancing the initial warming or cooling trend. An important positive feedback mechanism is related to the albedo of ice and is called the ***ice-albedo feedback***. When the Earth's climate cools, polar ice sheets expand and the surface of ice reflecting sunlight increases. Consequently, Earth's albedo increases and it further enhances the initial cooling. When Earth's climate warms, polar ice sheets retreat and the surface of ice reflecting sunlight decreases. Earth's albedo decreases and it further enhances the initial warming. Other feedback mechanisms probably involve ***greenhouse gases***. The remarkable correlation between global temperature and the atmospheric concentration of CO_2 and CH_4 (Fig. 75) emphasizes the sensitivity of Earth's climate to the greenhouse gases.

What initially triggered glacial-interglacial cycles is not fully understood. The oldest evidence of an ice sheet in Antarctica is 45.5 million years old (vs. less than 10 million years for the Greenland ice sheet). The initiation of the glaciations in the southern hemisphere coincides with a global cooling which started 50 million years ago (Fig. 76). One possible mechanism responsible for this global cooling is the uptake of atmospheric CO_2 by weathering of silicate rocks exposed during the formation of the Himalayas. The Himalayas began to form between 40 and 50 million years. Rocks caught between the Indian and Eurasian continents were uplifted and deformed, producing one of the largest mountain range on Earth. The process of mountain building increases considerably the surface of rocks exposed to weathering. The weathering of silicate rocks consumes atmospheric CO_2 (see chapter 9). Therefore, the formation of the Himalayas may have led to a net removal of atmospheric CO_2, contributing to the observed cooling trend of the past 50 million years.

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* Note that what is measured to determine this parameter is the ratio $^{18}O/^{16}O$ of the oxygen trapped in air bubbles in the ice, not the ratio $^{18}O/^{16}O$ of the ice (H_2O) itself. In section 10.3, we talk about the relationship between the ratio $^{18}O/^{16}O$ of the ice and ocean water and the extent of polar ice sheets. The ratio $^{18}O/^{16}O$ of the atmospheric oxygen is also related to the volume of polar ice but the explanation of this relationship does not only involve the hydrological cycle but also, and more importantly, biological processes. Although very interesting, this discussion lies beyond the scope of this course.

Figure 76: Evolution of the global deep ocean temperature over the past 60 million years. Note the cooling trend initiated 50 million years ago. Plio: Pliocene. Pt: Pleistocene. Source: adapted from Hansen et al. (2008).

8.2.2. Paleozoic and Proterozoic glaciations

Milankovitch cycles have probably affected the climate throughout Earth’s history. However, Earth was deprived of polar ice caps during most of its history. The glacial-interglacial cycles which characterized the climate of the last 2,000,000 years are a relatively unusual feature of Earth’s climate.

The absence of polar ice sheets is associated with high sea surface temperature and high levels of CO_2 and other greenhouse gases in the atmosphere. Consequently, the Earth devoid of polar ice caps is often referred to as the greenhouse Earth, as opposed to the icehouse Earth in which we live today[*].

The main driver of climate over millions of years is plate tectonics. Plate tectonics influences global climate on the long term through three main processes:

(1) Plate tectonics controls the presence or absence of a landmass at the poles. Snow and ice accumulate more readily on a landmass than on the sea surface.

(2) Plate tectonics controls the position of continents and the shape of ocean basins, which influence the atmospheric and ocean circulation, hence the way heat is carried from low to high latitudes.

(3) Plate tectonics controls the long-term trends in atmospheric CO_2 through (1) volcanism (input) and (2) weathering of silicate minerals (output).

8.3. How can scientists reconstruct the climate of the past?

There are many indicators of past climate (or ***climate proxies***). For instance, fossils of animals and plants can be used to reconstruct past climate because the composition of biological communities depends on climate conditions (e.g. arctic ecosystem vs. tropical ecosystem). The nature and structure of sediments is also influenced by climate. Ice for instance is associated to specific sedimentary structures that can be recognized in the geological record and used as evidence for the presence of past ice sheets.

Another mean of reconstructing past climate was used to draw the curve of figure 72. The curve shows the succession of glacial and interglacial periods over the past 1.8 million years. Maxima

[*] Note that we are experiencing a warm interglacial within the icehouse interval.

are correlated with warmer climates during which the extent of polar ice sheets was minimum. Minima are correlated with colder climates during which the extent of polar ice sheets was maximum. What kind of data was used to draw this curve?

To answer this question, we must turn our attention to the isotopes of oxygen. Isotopes of a given element have the same number of electrons (and protons) but have different numbers of neutrons. The oxygen has three stable isotopes: oxygen-16 (^{16}O), oxygen-17 (^{17}O), and oxygen-18 (^{18}O). The number associated to these isotopes is their mass number (= number of protons + number of neutrons). The higher the mass number, the heavier the isotope. In the present discussion, we are concerned with two isotopes only: the light ^{16}O and the heavy ^{18}O.

To understand the relationship between climate and these isotopes, one must first examine the behavior of these isotopes in the hydrological cycle (the water cycle). When seawater evaporates from the oceans, the water vapor (H_2O) is enriched in the light ^{16}O (Fig. 77). Some of the water that evaporates from the ocean condensates and returns to the ocean as rainwater. Rainwater is enriched in the heavy ^{18}O relative to the water vapor remaining in the atmosphere. A fraction of the water vapor ultimately reaches the poles and accumulates as ice and snow. Therefore, the ice and snow of polar ice caps are very much enriched in light ^{16}O relative to ocean water. During cold glacial periods (ice ages), polar ice caps are larger and therefore more of the light ^{16}O is trapped in polar ice. Consequently, less light ^{16}O remains in the oceans and the ratio $^{18}O/^{16}O$ of polar ice is very small whereas that of the ocean water is very large. Conversely, during warm interglacial periods, polar ice caps are smaller and more of the light ^{16}O is left in the oceans. The ratio $^{18}O/^{16}O$ of polar ice therefore increases during warm interglacial periods whereas that of the ocean water decreases (Fig. 77). In conclusion, the ratio $^{18}O/^{16}O$ of seawater and polar ice depends on the extent of polar ice sheets and is therefore an excellent paleoclimate indicator. Note also that mathematical formula can be used to deduce atmospheric and ocean water temperature from the ratio $^{18}O/^{16}O$.

But how can scientists reconstruct the evolution of the ratio $^{18}O/^{16}O$ of seawater and polar ice over millions of years? Scientists measure the isotopic ratio $^{18}O/^{16}O$ of the oxygen in the calcium carbonate ($CaCO_3$) of shells of marine organisms present in sediment cores and in the ice (H_2O) from cores drilled in the polar ice caps.

The longest climate record obtained from a polar ice core is 800,000 years. Ocean sediments can offer a much broader window on past climate changes dating back tens of millions of years. For instance, the curve of figure 73 was established by compiling data from many marine sediment cores to obtain a climate record spanning the past 5.3 million years (only part of the record is shown in figure 73).

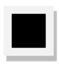

Figure 77: The ratio $^{18}O/^{16}O$ of water in relation with the hydrological cycle. The curve illustrates the evolution of the ratio $^{18}O/^{16}O$ of ocean water during warm interglacial periods (red) and cold glacial periods (blue). See text for explanation.

9. Recent global climate change

9.1. The rise of environmentalism in the 20th century

Human populations are profoundly affecting the environment, so much so that a new geologic epoch has been proposed in 2003: the ***Anthropocene***.

One number which perhaps reflects best the rising influence of humankind on the environment is the world's population. From 1 billion at the beginning of the 19th century, the world's population increased to 6 billions in 2000 and is expected to be between 8 to 11 billion by 2050 (projection of the Department of Economic and Social Affairs of the United Nations). We modify our landscape by constructing megacities. Our increasing energy needs still rely heavily on burning fossil fuels which releases greenhouse gases and modifies Earth's climate. A number of human activities are affecting the biosphere and causing species extinctions: the expansion of cities, deforestation for agricultural purposes, global warming, and the discharge of pollutants of all sorts in the environment. Understanding their effects on the environment is the prerequisite to taking measures to protect our environment.

Positive outcomes have emerged from our advancing knowledge of human interferences with the environment. For instance, in the 70s scientists raised public awareness on the destructive effect of chlorofluorocarbon (CFC) on the ozone layer, a layer of the atmosphere protecting the biosphere against UV radiations. The manufacturing of this compound is now strictly regulated in 197 states which have ratified the Montreal Protocol in 1987, an international treaty devised to protect the ozone layer. In parallel, ban on leaded gasoline began in the 70s after the American geochemist Clair Patterson[*] drew public attention on the abnormally high concentrations of lead in the environment.

In 1988, the Intergovernmental Panel on Climate Change (IPCC) was established by the United Nations. The IPCC consists of a group of climate experts whose role is to inform the public and governments on the impact of human activities on climate, on predictions of future climate change, and on solutions to mitigate human-induced warming. They publish reports based on the work of thousands of scientists from around the world. In 1997, an international treaty called the Kyoto Protocol was designed to reduce the global emission of greenhouse gases. It has been ratified by 191 countries and the European Union. Thirty seven industrialized countries and the European Union are legally bound by the treaty to reduce their emission of greenhouse gases. The United States have not yet ratified the treaty and Canada pulled out in 2012.

Other efforts that are being made to protect the environment involve the designation of protected areas. For instance, the Great Barrier Reef Marine Park was established in 1975 by the Australian government. The park covers 344,400 km^2. Tourism is strictly regulated and fishing activities are banned in one-third of the park.

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[*] The same geochemist who determined the age of the Earth based on the age of meteorites using the uranium-lead dating technique, a technique he himself developed. He found an age of $4.55 \pm 0.07 \times 10^9$ years (published in 1956) which is still the accepted age of the Earth today.

9.2. Evidence for recent global climate change

Reliable instrumental measurements of atmospheric temperatures are available since the end of the 19th century. These measurements show that the global average temperature has increased by about 0.8°C over the past century. As we saw in chapter 10 paleoclimate indicators can be used to reconstruct the evolution of Earth's climate in the past. An important finding is that Earth has already been as warm and even warmer during the course of its history. For example, the early Eocene epoch (50 million years ago) was warmer than today with no ice at the poles and a rainforest flourishing in the arctic region! During the past 10,000 years (Holocene) the Earth's climate has been relatively stable (see fig. 89-B). The rate at which global temperature is now rising is exceptional compared with the climate record of the past 1000 years (Fig. 78).

Figure 78: Evolution of the average temperature of the northern hemisphere relative to the average temperature of the period 1961-1990. Source: IPCC 2007 Assessment Report.

The rapid increase in global temperature correlates with the rise of atmospheric greenhouse gases caused by human activities since the industrial revolution which took place during the late 18th and early 19th centuries (Fig. 79). Moreover, the geological record shows that current levels of atmospheric CO_2 and CH_4 are unprecedented during the past 650,000 years (Fig. 80). The atmospheric concentrations of CO_2 and CH_4 have increased by 40% and 250%, respectively, since pre-industrial times. Another important greenhouse gas produced by human activities is N_2O (nitrous oxide).

Figure 79: Atmospheric concentrations of CO_2 and CH_4 over the last 10,000 years. Levels prior to reliable instrumental measurements of atmospheric samples are from ice cores. Source: IPCC 2007 Assessment Report.

Figure 80: Evolution of atmospheric concentration of three greenhouse gases over the past 650,000 years measured in ice cores from Antarctica and, for recent concentrations, in atmospheric samples. Stars in the upper right corner indicate the concentrations in 2000. Source: IPCC 2007 Assessment Report.

Complex computer models are used to understand how the climate system responds to various perturbations. These models show that the current global warming cannot be reproduced accurately without the contribution of anthropogenic greenhouse gases. Natural phenomena alone cannot explain the recent rise of global temperature (Fig. 81). The cooling effect of aerosols, which reflect the incoming sunlight, and solar activity, which has been relatively low in recent years, have not offset the global warming that has been detected over the past decades (fig. 82)

These models are also used to predict the future course of Earth's climate. The results greatly depend on the evolution of atmospheric greenhouse gases. The IPCC reports propose several scenarios (Fig. 83), from a pessimistic scenario with almost no reduction in emission of greenhouse gases (scenario RCP8.5) to a very optimistic scenario predicting a generalized use of non-fossil energy sources and active removal of carbon dioxide (scenario RCP2.6).

Figure 81: Comparison between observed changes in surface temperature (black line) and those simulated by climate models ("natural forcings only" models in blue and "natural forcings + anthropogenic forcings" models in pink). Source: IPCC 2013 Assessment Report.

Figure 82: Principal components of the radiative forcing of climate change. Radiative forcing is defined as the difference between the solar energy reaching the Earth (incoming sunlight) and the energy radiated back to space. A positive radiative energy (energy absorbed by Earth) causes warming, whereas a negative radiative energy (energy lost to space) causes cooling. Source: IPCC 2013 Assessment Report
(via http://www.epa.gov/climatechange/science/indicators/ghg/climate-forcing.html).

Figure 83: Evolution of the average global temperature in the future predicted by models based on different scenarios of CO_2 emissions. Source: IPCC 2013 Assessment Report.

9.3. Consequences of global climate change

I. Impact on regional weather patterns

The consequences of global warming on regional weather patterns are complex. The term global warming does not mean that all regions of the world are experiencing the same amount of warming (or even any warming at all!). The trends are not uniform. Figure 84 shows the differences between the average surface temperatures measured during the period 2000-2010 and the average temperatures of the period 1940-1980 measured in the same region.

Figure 84: Differences in annual mean temperatures measured during the period 2000-2010 and those measured during 1940-1980 in the same region. Source: NASA (http://data.giss.nasa.gov/gistemp/maps/).

The figure above clearly shows that the warming of the northern hemisphere is more pronounced than the warming in the southern hemisphere. Moreover, warming is more pronounced over the land than over the ocean. The evolution of surface temperature will therefore vary geographically.
Global warming is also disrupting precipitation patterns (Fig. 85).

Figure 85: Changes in precipitation for the period 1900-2013. Source: Mark Maslin (Climate Change, 2014).

Presumably global warming may also affect the frequency of intense climatic events such as powerful storms, cyclones/typhoons and heatwaves. Models predict that global warming will increase the number and intensity of storms and cyclones. However, no clear link between global warming and recent trends in the intensity and frequency of storms and tropical cyclones has been established. Longer-term monitoring of atmospheric and oceanographic conditions is needed to understand the influence of global warming on the life cycle of these violent climatic events.

II. Impact on the cryosphere

Global warming accelerates the melting of polar ice sheets. Whereas there is no doubt that the Greenland ice sheet is rapidly shrinking, there has been some controversy surrounding the fate of the Antarctic ice sheet. Recent satellite data confirm that both the Greenland and Antarctic ice sheets are loosing ice (Fig. 86).

Figure 86: Changes in ice mass estimated from data collected by the GRACE satellite for the Greenland ice sheet (A) and the Antarctic ice sheet (B) . Source: Velicogna (2009).

A direct consequence of the melting of the ice sheets is sea level rise (Fig. 87 on this page). Another cause of sea level rise related to global warming is the increasing volume of the oceans due to warming ocean waters, i.e. ***thermal expansion*** of the oceans. Sea level rise will increase the frequency of storm surges in low-lying coastal areas. Areas at risk include the Netherlands, Bangladesh and New York. Many low-lying South Pacific islands are also directly threatened by sea level rise, like the Maldives and Tuvalu (see Fig. 87 on page 86). The sea level is currently rising at a rate of 2-3 mm/year. Based on predictive models, the sea level is expected to rise at a rate of 2-6 mm/year during the next 100 years.

The rapid warming over high-latitude landmasses in the Northern Hemisphere may also result in a loss of permafrost that can further enhance the initial warming (see section 11.3.**IV**.).

Figure 87: On the previous page, future evolution of global mean sea level based on various scenarios of CO_2 emissions (sources: IPCC 2013 Assessment Report). Above, picture of the island of Malé, capital of the Maldives, showing how vulnerable some low-lying Pacific islands are to sea level rise (photo by Hironobu Kan, Kyushu University).

III. Impact on the biosphere

We have already mentioned the impact of ***ocean acidification*** on marine ecosystems related to the increase in atmospheric CO_2. Ocean acidification may have a detrimental effect on calcifying organisms like corals. Global warming itself may be detrimental to corals by increasing the frequency and intensity of ***coral bleaching*** events[*].

Global warming also affects the geographic distribution of organisms. Organisms are expected to migrate poleward or to higher altitudes to escape the heat. ***Poleward migration*** has already been observed in several groups of organisms such as corals, butterflies, and birds. The introduction of new species in areas where they were previously absent will affect species interactions. Introduction of ***new competitors*** may threaten the survival of certain species. Organisms which have a restricted climatic tolerance and cannot migrate (or not fast enough) will also be at risk of extinction. Figure 84 shows that one of the geographic areas that is the most affected by global warming is the Arctic region. Reduction in sea ice volume due to global warming leads to ***reduction of habitat*** and potential threat of extinction for organisms living on the ice, like polar bears, walruses and seals (Fig. 88A-B). On the other hand, there are also species that will benefit from global warming. A recent study reports an increase in the population of Adélie penguins on Beaufort Island due to an increase in breeding habitat related to the retreat of glaciers (Fig. 88C).

Another risk of global warming is the spread of ***contagious tropical diseases*** to high-latitudes, for example, malaria which is transmitted to humans by mosquitoes thriving in tropical to subtropical regions.

In conclusion, global warming could cause irreversible changes in the biosphere. The IPCC estimates that 20 to 30% of known species of animals and plants may face extinction if global average temperature increases by 1.5 to 2.5°C.

Figure 88: Loosers and winners of global warming. (A) Polar bear, (B) walruses (source of A and B: Greenpeace Technical Report 04-2012), (C) satellite image of Antarctica and Beaufort Island where the population of Adélie penguins has expanded (source: LaRue et al., 2013).

IV. Abrupt environmental changes

So far we have considered a steady rise in global temperature and progressive changes in weather patterns, the cryosphere and the biosphere. However, we know from the geological record that these changes are not always progressive. For instance, if we look at the sea level rise of the last deglaciation (Fig. 89), we see that there is a big jump around 14,500 years ago (MWP-1A in Fig. 89). It is called a ***melt water pulse***. The sea level rose by about 14-18 m in 350 years. The rate of sea level rise exceeded 40 mm/year which is more than 10 times faster than the current sea

[*]Corals are marine invertebrates containing photosynthetic microalgae inside their living tissue. Corals and microalgae have a symbiotic relationship, meaning that they mutually benefit from their association. Corals provide microalgae with a protected shelter. Through cellular respiration, corals also provide the CO_2 needed to perform photosynthesis. Microalgae provide corals with organic compounds produced by photosynthesis. This remarkable relationship explains why corals need light to thrive and tropical coral reefs grow in shallow waters. Corals and their symbiotic microalgae are sensitive to temperature. If the ocean water warms too much, corals loose their microalgae. Since corals get their color from their microalgae, loosing them means they loose their color and become white, hence the term coral bleaching. If the water temperature does not return to normal soon enough, corals do not regain their microalgae and perish.

level rise! Scientists relate this jump in sea level to a sudden collapse of the polar ice sheets, either in the northern or southern hemisphere. In the present context of accelerated melting of polar ice, it is important to understand the mechanism triggering these events to be able to predict future sea level jumps.

Figure 89: (A) Sea level change during the last deglaciation. Note the sea level jump (MWP-1A) around 14,500 years ago. (B) Evolution of atmospheric temperature in the Arctic during the same period. Superimposed on this curve is the evolution of average summer insolation at 65°N. Source: Montaggioni and Braithwaite (2009).

An event related to global warming that could trigger rapid and intense climatic changes is a ***slowdown of the thermohaline circulation***. As we learnt in chapter 8, the thermohaline circulation is driven by differences in water density. An important source of deep water is the North Atlantic where cold and salty —hence dense— water sinks to the bottom of the ocean. The thermohaline circulation helps redistribute heat from the tropics to higher latitudes in the North Atlantic via the Gulf Stream. The discharge of a large amount of freshwater from melting ice and increased precipitation may disrupt the thermohaline circulation by lowering the density of surface water. This could potentially affect regional weather patterns, particularly in Europe.

We have already seen that acceleration and intensification of a preexisting climatic trend, be it warming or cooling, may be caused by positive feedback mechanisms. One such positive feedback that could intensify the current warming is related to permafrost thawing (Fig. 90). Melting of the permafrost could result in a rapid decay of the organic matter it contains. The microbial decomposition of organic matter would release large amount of carbon dioxide and methane, two greenhouse gases. Methane is a particularly potent greenhouse gas. Another source of methane may come from the deep sea. A few hundreds meters beneath the sea floor, methane is sometimes present as methane hydrate (or clathrate). Methane hydrate is formed of water ice containing molecules of methane trapped in the crystal lattice of ice (Fig. 91). The methane originates from microbial activity occurring in deep-sea sediments. Methane hydrate is stable at relatively high pressure and low temperature and occurs between 200 and 400 m beneath the sea floor. Global warming could potentially destabilize methane hydrate, which would lead to a massive release of gaseous methane in the atmosphere. Release of methane in the atmosphere through ***permafrost thawing*** and/or ***destabilization of methane hydrate*** would intensify global warming. The resulting positive feedback loop may result in more methane released in the atmosphere causing a rapid and intense warming. Evidence suggests that such an event may have already happened in the geological past and even caused mass extinctions.

Figure 90: Distribution of permafrost in the northern hemisphere. Source: International Permafrost Association

Figure 91: Left : structure of methane (source : NASA). Middle and right : methane hydrate (sources: AP images and NASA).

V. Socioeconomic impacts

The socioeconomic impact of global warming will depend on the geographic area. Some regions will be more negatively affected than others. Some may even benefit from a moderate warming, e.g. wineries in the UK. However, scientists have defined a limit of global warming beyond which the impact would be negative for most people on Earth. That limit has been set to a 2°C increase above pre-industrial temperature.

Climate change may affect human societies through its impact on a broad range of domains:

- Freshwater resources (groundwater recharge influenced by droughts)
- Food security (crop yield directly influenced by climate)
- Coastal systems (sea level rise and storms increase the rate of coastal erosion and the risk of floods in coastal areas)
- Ecosystems (e.g. the decline of coral reef ecosystems affects people for whom coral reefs are a source of food and revenue from tourism)
- Human health (heatwaves, droughts, storms, floods and contagious diseases directly affect human health)

9.4. Strategies to deal with anthropogenic global climate change

One can think of several possible strategies to deal with climate change. One of them is simply adaptation, for example, protect coastal areas against floods or switch to other types of crops requiring less water. Another solution is to reduce the emission of greenhouse gases and try to mitigate the current global warming (mitigation strategy). The Kyoto Protocol was designed for this purpose. Although some significant effort has been made, countries with fast-growing economies like China or India release more CO_2 in the atmosphere every year (Fig. 92). Efforts are being made to devise cleaner sources of energy (see next chapter). Energy-saving plans can also be adopted, for example, by improving building insulation or increasing the energy efficiency of lighting devices (e.g. LED). In addition, economic strategies to cut CO_2 emissions have also been devised: carbon taxes and carbon trading. The former consists simply of a tax that has to be paid by industries producing CO_2. The latter consists of putting a price tag on CO_2 emissions so that a country that emits more is able to purchase the right to emit to a country that emits less. The goal is to force countries to respect their engagement in cutting CO_2 emissions.

Figure 92: Evolution of anthropogenic emission of CO_2 in Japan (A), China (B), Germany (C) and India (D). Source: Carbon Dioxide Information Analysis Center.

Another set of strategies are called geoengineering strategies. Geoengineering means interfering with natural processes on a large scale with the purpose of making the environment more favorable for human populations. An example of geoengineering technique is triggering precipitation by releasing in the atmosphere chemicals which promote the formation of rain droplets or ice particles (i.e. cloud seeding). Geoengineering actions have also been devised to slow down or halt the current global warming (Fig. 93). Geoengineering techniques can be classified into two categories:

- Controlling the amount of solar radiation absorbed by Earth's surface
- Actively removing carbon dioxide from the atmosphere.

Figure 93: Examples of geoengineering strategies. Note that if carried out on a planetary scale, reforestation and carbon sequestration can also be called geoengineering. Source: Keith (2001).

Some examples of geoengineering strategies are described below:

Reforestation

Sustainable management of existing forests and planting new trees are solutions which can slow the increase of atmospheric carbon dioxide based on the ability of plants to store carbon through photosynthesis. This solution also includes the potential use of genetically modified plants that could fix carbon with a greater efficiency.

Carbon dioxide capture and storage

Another solution consists in injecting the carbon dioxide produced by industries into underground geological reservoirs. This is called ***carbon sequestration*** or carbon dioxide capture and storage (Fig. 94).

Figure 94: Carbon sequestration. Source: IPCC 2005 report "Carbon Dioxide Capture and Storage".

Iron fertilization

Iron is essential for the growth of phytoplankton (photosynthetic microaglae) which is often naturally limited by the low abundance of this element in seawater. Hence, dumping iron in the ocean can boost primary production. More primary production means more carbon fixed by photosynthesis leading to a net removal of atmospheric CO_2.

Albedo modification

The release of aerosols (tiny droplets) in the upper atmosphere and the installation of giant mirrors in space are ideas put forward by some scientists as possible strategies to block part of the incoming sunlight and induce a global cooling which would oppose the current warming trend.

12. Geological resources

See slides