## 9. The age of rocks

### 9.1. Absolute vs. relative ages

The issue of time is central to earth science. Geologists want to know how old rocks are and how long it takes for geological processes to occur. Until the 20th century and the discovery of radioactivity, geologists could only tell whether a rock was younger or older than another rock. In other words, they could only determine the ***relative age*** of rocks. Knowing the number of years elapsed since a rock has formed, in other words knowing the ***absolute age*** of the rock, could only be achieved after the development of radiometric dating in the 20th century.

### 9.2. The relative age of rocks

*9.2.1. Stratigraphic principles*

***Stratigraphy*** is the study of sedimentary layers (or strata). The founder of stratigraphy is a Danish scientist named Nicolas Steno (1638-1686). He is also known for his study of the fossil *Glossopetrae* which he correctly identified as fossilized shark teeth, a notable achievement at a time when the true nature of fossils was poorly understood. His interest for geology led him to study the sedimentary rocks of northern Italy (where he lived) and to enounce two fundamental rules of stratigraphy:

1. **Principle of original horizontality**

Sediments are deposited horizontally (or nearly so) by gravity. This implies that sedimentary layers which are folded or faulted have been subjected to deformations after their deposition.

2. **Principle of superposition**

New sedimentary layers form on top of older layers. Therefore in an undeformed succession of sedimentary layers, layers at the bottom are older than layers at the top.

The principle of superposition is of course very useful to determine the relative age of sedimentary layers. When studying a section of sedimentary rocks, this principle can be used to reconstruct the chronological order of the layers providing that we understand the deformation history of the rocks in question[*].

What if we have two sections that are very distant from each other? How can we know that the rocks of one section are younger or older than the rocks of another section located far away? To solve this problem, one way is to study the fossil content of sedimentary layers.

*9.2.2. Biostratigraphy*

[*] Due to folding, a succession of sedimentary layers can become up-side down and the chronological order inverted. If the deformation is not noticed, the chronological reconstruction will be completely wrong!

***Biostratigraphy*** is the branch of stratigraphy which uses fossils to correlate sedimentary rocks and find out their relative ages. The technique relies on the fact that life evolves. Some species go extinct while new species appear. As a consequence, fossil assemblages of different geological intervals are not the same. The geological record is characterized by fossil assemblages whose species composition is changing through time. This is the ***principle of faunal succession***. For example, a fossil of Australopithecus (one ancient genus of hominids —the family of primate to which we belong— who appeared around 4 million years ago and went extinct 2 million years later) cannot be found in the same layer as a fossil of dinosaur because dinosaurs went extinct 65 million years ago (61 million years before the first Australopithecus appeared!).
Some species are more useful than others when it comes to determine the relative ages of sedimentary rocks. Widespread species which have existed for a limited time and are characteristic of specific geological periods are excellent biostratigraphic markers. These species are called ***index species***.

**NB**: the concept of the evolution of life was accepted before Charles Darwin (1809-1882) proposed his theory of evolution by natural selection. The roots of biostratigraphy are to be found in the 18$^{th}$ century when scientists noticed that sedimentary rocks of different ages were characterized by different fossils. What Darwin discovered is a mechanism explaining how new species arise and therefore how evolution can take place.

### *9.2.3. Unconformities: gaps in the stratigraphic record*

The stratigraphic record is not continuous. There are gaps which can result from prolonged lack of deposition or from the erosion of preexisting sedimentary layers. The surfaces corresponding to these gaps are called ***unconformities***.
In many cases unconformities result from an episode of erosion. Sedimentary layers formed in a lake or in the sea can be eroded during a long period of emergence. How do sedimentary layers initially formed under water become emerged? There are two important mechanisms that can cause long period of emergence: ***sea level fall*** **(1)** and upward ground movement or ***uplift*** **(2)**.

**(1)** During ice ages (glacial periods), more ice accumulates on landmasses and sea level falls dramatically. During the maximum of the last ice age about 20,000 years ago (Last Glacial Maximum = LGM), the sea level was 125 m lower than today! The continental shelves stood above sea level and sediments were exposed to the erosional action of the wind and the rain. New sedimentary layers formed only when the sea re-flooded the continental shelves during the current warm period. The surface which separates these new layers from the layers deposited before the LGM is an unconformity and represents a time gap of thousands of years.
**(2)** Regions of the crust near convergent plate boundaries experience tremendous compressional forces which can cause folding and bring portions of the crust at higher elevations, a process

called tectonic uplift*. Uplifted sedimentary layers can be raised above sea level where they are exposed to erosional processes. Once compression stops, an episode of subsidence (downward movement of the crust, see section 7.5) may create a depression in which new sediments can accumulate. The surface between the older folded sedimentary layers and the new sediments is an unconformity which may encompass a time lapse of millions of years.

There are three types of unconformities:

***Disconformity***: the sedimentary layers below and above the unconformity are both undeformed and horizontal. An example of disconformity is the unconformity corresponding to the episode of sea level fall presented above [see **(1)**].

***Angular unconformity***: the sedimentary layers below the unconformity are folded whereas the sedimentary layers above the unconformity are undeformed and horizontal. An example of angular unconformity is the unconformity related to the episode of compressional deformation followed by erosion, subsidence, and deposition of new sediments presented above [see **(2)**].

***Nonconformity***: unconformity between sedimentary rocks and metamorphic or igneous rocks. This type of unconformity may represent a very extensive time gap of tens of millions of years or more.

*9.2.4. Cross-cutting relationships*

The geometrical relationships between sedimentary rock formations, igneous intrusions (e.g. dykes —sheet-like intrusion intersecting rock layers—), faults, and unconformities can be used to reconstruct the chronological order in which these geological features formed. The rule is simple: *younger geological structures cut older ones*. For example, a dyke intersecting sedimentary layers must be younger than these layers or a fault cutting a dyke must be younger than the dyke.

*9.2.5. The geological time scale*

Geologists of the 19th century used the principle of superposition and biostratigraphy to divide the geological record into successive intervals characterized by distinct fossil assemblages. This approach led to the construction of the ***geologic time scale***. The basic subdivisions of the geologic time scale (from longer to shorter time units) are ***eras*** (e.g. Cenozoic), ***periods*** (e.g. Quaternary), and ***epochs*** (e.g. the current epoch called Holocene).

The boundary between geologic periods is characterized by abrupt changes in fossil assemblages. Several of these boundaries correspond to mass extinctions. A mass extinction event represents a relatively short geologic time span (a few million years or less) during which a large proportion of the total number of species living on the Earth (e.g. 75% or more) disappears (see chapter on the evolution of life).

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* Note that uplift does not necessarily imply folding. Rocks can be uplifted without undergoing much folding.

## 9.3. The absolute age of rocks

When the geologic time scale was established in the $19^{th}$ century, geologists did not know the duration of each period. Nobody knew precisely how old the Earth was. This issue was important not only for science but also for religion because of the strongly held religious belief that the Earth could not be older than a few thousand years based on the rigorous interpretation of religious texts. The solution came only after the discovery of radioactivity in 1896 by a French physicist named Henri Becquerel. At the beginning of the $20^{th}$ century, the physicist Ernest Rutherford proposed a technique to date rocks based on radioactive decay. The method of ***radiometric dating*** was born. Based on this method, an American geochemist named Clair C. Patterson calculated in the 1950s an age for the Earth of 4.56 billion years!

The basic principle of radiometric dating is simple. Elements consist of different isotopes with nuclei composed of the same number of protons but with different numbers of neutrons. ***Radioactive isotopes*** are isotopes which spontaneously disintegrate (decay) into a different element. In a rock containing a certain amount of a given radioactive isotope, the concentration of the element produced by radioactive decay increases with time and can be used as a natural clock to determine the age of the rock. This can work only if the rate of radioactive decay is known and does not vary in time.

The rate at which a radioactive isotope (parent atom) disintegrates into another element (daughter atom) is expressed by its ***half-life***. The half-life of a radioactive isotope is the time its takes for half of the initial amount to decay into daughter atoms (Figs. 1 & 2).

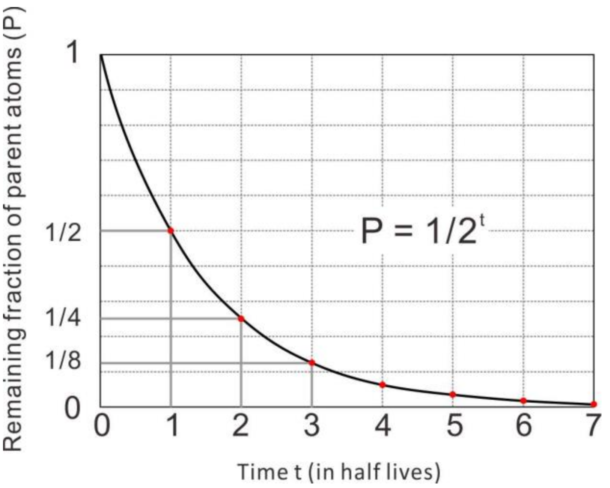



Figure 1

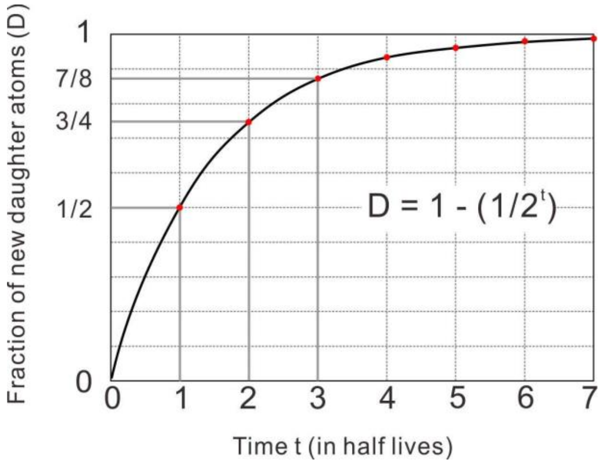



Figure 2

Different isotopes have different half-lives. For example, $^{14}C$ (carbon-14) disintegrates into $^{14}N$ and has a half life of 5730 years. Another example is $^{87}Rb$ which disintegrates into $^{87}Sr$ and has a half-life of 49 billion years. Therefore if you want to measure the age of rocks that are hundreds of millions of years old or more, you can use the rubidium-strontium system but not the radioactive isotope of carbon. The $^{14}C$ method is used to measure the age of much younger materials. For example, you can use $^{14}C$ to measure the age of corals which were growing thousands of years ago.

*The half-life of a particular radioactive isotope is constant*. It is not affected by changes in physic-chemical conditions, such as variations in temperature and pressure. This is of fundamental importance because if the half-life of radioactive isotopes varied in time, radiometric dating would not be possible.

In order to understand how the age of a rock can be measured, let's take the example of the rubidium-strontium system. In this case, the parent atom is $^{87}Rb$ and the daughter atom is $^{87}Sr$. Let's say we collected a sample of igneous rock and we want to know its age. The age of the sample in this case is the time elapsed since it crystallized from a cooling magma. The crystals forming in the cooling magma have trapped a certain amount of parent atoms $[^{87}Rb]_{t=0}$. First let's suppose that there is no daughter atoms $^{87}Sr$ incorporated in the crystals when they formed (unrealistic!). We can express the amount of $^{87}Rb$ and $^{87}Sr$ present in our sample as a function of time using the following relationship:

$$[^{87}Sr]_t = [^{87}Rb]_{t=0}\ [1 - (1/2^t)]$$

$$[^{87}Rb]_t = [^{87}Rb]_{t=0}\ (1/2^t)$$

$$\frac{[^{87}Sr]_t}{[^{87}Rb]_t} = \frac{[^{87}\cancel{Rb}]_{t=0}}{[^{87}\cancel{Rb}]_{t=0}}\ \frac{[1 - (1/2^t)]}{1/2^t} = \frac{[1 - (1/2^t)]}{1/2^t} = \frac{1}{1/2^t} - \frac{1/2^t}{1/2^t} = 2^t - 1$$

$$\boxed{[^{87}Sr]_t = (2^t - 1)\ [^{87}Rb]_t} \qquad \text{(A)}$$

Equation (A) represents a straight line with a slope equal to ($2^t$-1) (Fig. 3). Therefore measuring the amount of $^{87}Sr$ and $^{87}Rb$ in our sample would enable us to calculate the age of the rock.

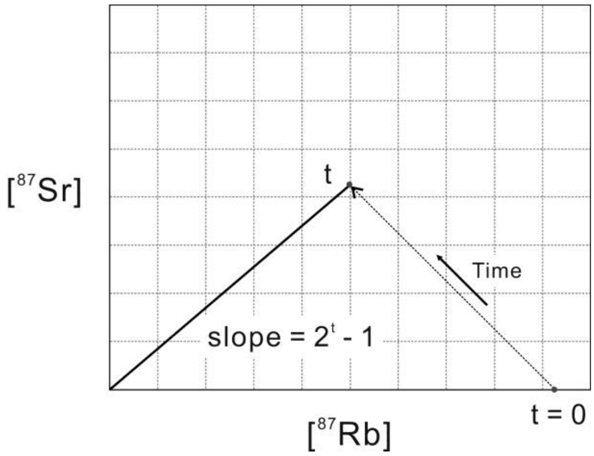



Figure 3

In reality, there is always a certain amount of daughter atoms $^{87}Sr$ which is incorporated in the crystals when the rock forms. In such case, equation (A) becomes (Fig. 4):

$$\boxed{[^{87}Sr]_t = (2^t - 1)\ [^{87}Rb]_t + [^{87}Sr]_{t=0}} \qquad (B)$$

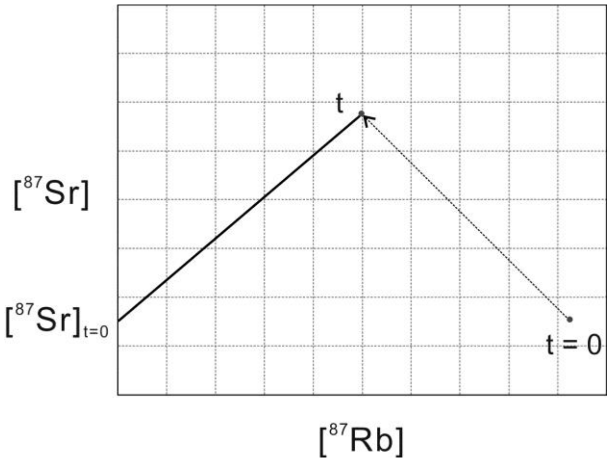



Figure 4

In this case, measuring the amount of parent atoms and daughter atoms in our sample is not enough to calculate the age of the rock because we don't know how much daughter atoms has been trapped initially in the rock (Fig. 5). Each minerals of our rock sample can incorporate any amount of $^{87}Rb$ and $^{87}Sr$ at the time of crystallization.

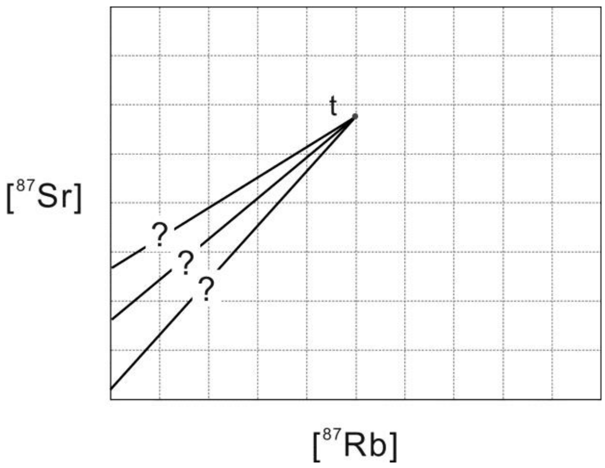
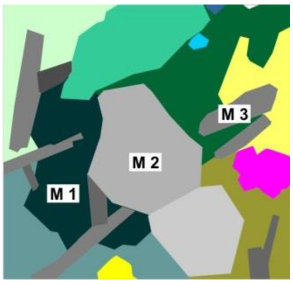



Figure 5

The solution to solve this problem is to consider the ratio of $^{87}Sr$ to a stable isotope of the same element which has the same properties. Here the stable isotope in question is $^{86}Sr$. Since $^{87}Sr$ and $^{86}Sr$ have the same behavior during physico-chemical reactions, they will incorporate different minerals always in the same proportion*. Let's say we have 1000 atoms of $^{87}Sr$ and 1200 atoms of $^{86}Sr$ initially present in the magma. The table below shows their distribution in three different minerals. The initial ratio $^{87}Sr/^{86}Sr$ is independent of the amount of $^{87}Sr$ trapped in the minerals when they crystallized.

| | Mineral 1 (M1) | Mineral 2 (M2) | Mineral 3 (M3) |
|---|---|---|---|
| $[^{87}Sr]_{t=0}$ | 500 | 100 | 400 |
| $[^{86}Sr]_{t=0}$ | 600 | 120 | 480 |
| $\left(\frac{[^{87}Sr]}{[^{86}Sr]}\right)_{t=0}$ | 0.83 | 0.83 | 0.83 |




If we divide equation (B) by $[^{86}Sr]_t$, we obtain the following relationship:

$$\boxed{\left(\frac{[^{87}Sr]}{[^{86}Sr]}\right)_t = (2^t - 1)\left(\frac{[^{87}Rb]}{[^{86}Sr]}\right)_t + \left(\frac{[^{87}Sr]}{[^{86}Sr]}\right)_{t=0}} \qquad (C)$$

We can thus determine the age of our sample by measuring the ratio of $^{87}Sr$ and $^{87}Rb$ to $^{86}Sr$ in several minerals in order to obtain a straight line from which we can derive the age t (Fig. 6). In

* Note that the isotopic composition of the magma should be uniform when crystallization occurs.

addition to the age, we can also find out the initial ratio of $^{87}Sr$ to $^{86}Sr$.

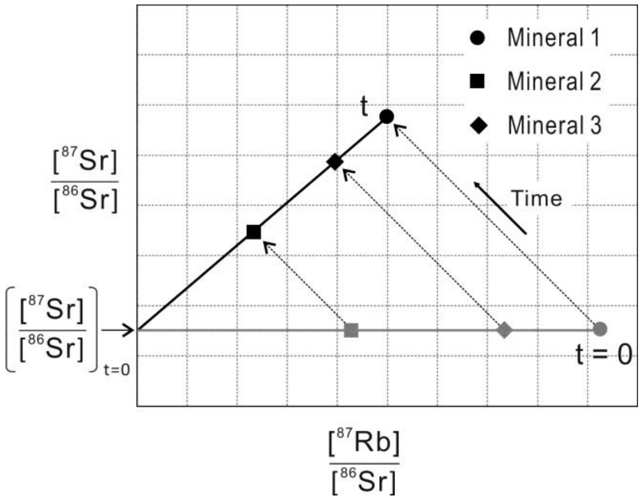



Figure 6

The slope of the line defined by the three minerals in the graph above is a function of the time elapsed since crystallization. This method is valid only if the minerals have not exchanged elements with the surrounding environment after they crystallized. Each mineral must have remained a ***closed system****. For metamorphic rocks, the age obtained corresponds to the time since the last phase of crystallization. We can also measure the age of biominerals like coral skeletons or algal crusts made of $CaCO_3$. In this case, we measure the time elapsed since the organism secreted the mineral assuming that the system remained closed afterward.

* Note that there are techniques dealing with open systems.