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.

[*] 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).

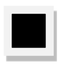

Figure 77: The ratio $^{18}O/^{16}O$ of water in relation with the hydrological cycle. The curve illustrates the evolution of the ratio $^{18}O/^{16}O$ of ocean water during warm interglacial periods (red) and cold glacial periods (blue). See text for explanation.