15.4: Prehistoric Climate Change

Figure \(PageIndex(1)\): Maximum extent of Laurentide Ice Sheet
Over Earth history, the climate has changed a lot. For example, during the Mesozoic Era, the Age of Dinosaurs, the climate was much warmer and carbon dioxide was abundant in the atmosphere. However, throughout the Cenozoic Era (65 Million years ago to today), the climate has been gradually cooling. This section summarizes some of these major past climate changes.

15.3.1: Past Glaciations

Through geologic history, the climate has changed slowly over millions of years. Before the most recent Pliocene-Quaternary glaciation, there were three other major glaciations [20]. The oldest, known as the Huronian, occurred toward the end of the Archean-early Proterozoic (~2.5 billion years ago). The major event of that time, the great oxygenation event (Chapter 8), is most commonly associated with the cause of that glaciation. The increased oxygen is thought to have reacted with the potent greenhouse gas methane, causing cooling [21].

The end of the Proterozoic (about 700 million years ago) had another glaciation, known as the Snowball Earth hypothesis [22]. Glacial evidence has been interpreted in widespread rock sequences globally and even has been linked to low-latitude glaciation [23]. Limestone rock (usually formed in tropical marine environments) and glacial deposits (usually formed in cold climates) are often found together from this time in regions all around the world. In Utah, Antelope Island in the Great Salt Lake has interbedded limestone and glacial deposits (diamictites) interpreted to be formed by continental glaciation [24]. The idea of the controversial Snowball Earth hypothesis is that a runaway albedo effect (ice and snow reflecting solar radiation) might cause the complete freezing of land and ocean surfaces and a collapse of biological activity. The ice-covered earth would only melt when carbon dioxide from volcanoes reached high concentrations, due to the inability for carbon dioxide to enter the then-frozen ocean. Some studies estimated carbon dioxide was 350 times higher than today’s concentrations [22]. The complete freezing [25] and the extent of the freezing [26] has come into question.

Glaciation also occurred in the Paleozoic, most notably with the Karoo Glaciation of the Pennsylvanian (323 to 300 million years ago). This also was caused by an increase of oxygen and a subsequent drop in carbon dioxide, most likely produced by the evolution and rise of land plants [27].

![Global Surface Temperature](https://example.com/figure.png)

**Figure (PageIndex[1]):** Global average surface temperature over the past 70 million years.

During the Cenozoic Era (the last 65 million years), the climate started out warm and gradually cooled to today. This warm time is called the **Paleocene-Eocene Thermal Maximum** and Antarctica and Greenland were ice-free during this time. Since the Eocene, tectonic events during the Cenozoic caused persistent and significant planetary cooling. For example, the collision of the Indian Plate with the Asian Plate created the Himalaya Mountains increasing weathering and erosion rates. An increased rate of weathering of silicate minerals, especially feldspar, consumes carbon dioxide from the atmosphere and
therefore reduces the greenhouse effect, resulting in long-term cooling [28].

![Diagram of the Antarctic Circumpolar Current](image)

**Figure 1:** The Antarctic Circumpolar Current

At about 40 Ma, the narrow gap between the South American Plate and the Antarctica Plate widened, resulting in the opening of the Drake Passage. This allowed for the unrestricted west-to-east flow of water around Antarctica, the Antarctic Circumpolar Current, which effectively isolated the southern ocean from the warmer waters of the Pacific, Atlantic, and Indian Oceans. The region cooled significantly, and by 35-million-year ago (Oligocene) glaciers had started to form on Antarctica [29].

At around 15 Ma, subduction-related volcanism between Central and South America created the Isthmus of Panama that connected North and South America. This prevented water from flowing between the Pacific and Atlantic Oceans and reduced heat transfer from the tropics to the poles. This created a cooler Antarctica and larger Antarctic glaciers. The expansion of that ice sheet (on land and water) increased Earth’s reflectivity (albedo), a positive feedback loop of further cooling: more reflective glacial ice, more cooling, more ice, and so on [30; 31].

By 5 million years ago (Pliocene Epoch), ice sheets had started to grow in North America and northern Europe. The most intense part of the current glaciation is the last 1 million years of the Pleistocene Epoch. The Pleistocene has significant temperature variations (through a range of almost 10°C) on time scales of 40,000 to 100,000 years, and corresponding expansion and contraction of ice sheets. These variations are attributed to subtle changes in Earth’s orbital parameters called **Milankovitch cycles** [32; 33], which are explained in more detail in the chapter on glaciers. Over the past million years, the glaciation cycles have been approximately every 100,000 years [34] with many glacial advances in the last 2 million years (Lisiecki and Raymo, 2005) [35].
Warmer portions of climate within an ice age are called **interglacials**, with brief versions called **interstadials**. These warming upticks are related to variations in Earth’s climate like Milankovitch cycles. In the last 500,000 years, there have been 5 or 6 interglacials, with the most recent belonging to our current time, the Holocene.

Two of the more recent climate swings demonstrate the complexity of the changes: the Younger Dryas and the Holocene Climatic Optimum. These events are more recent and yet have conflicting information. The Younger Dryas cooling is widely recognized in the Northern Hemisphere [36], though the timing of the event (about 12,000 years ago) does not appear to be equal everywhere [37]. It also is difficult to find in the Southern Hemisphere [38]. The Holocene Climatic Optimum is the warming around 6,000 years ago [39], though it was not universally warmer, and probably not as warm as current warming [40], and not at the same time everywhere [41].

### 15.3.2: Proxy Indicators of Past Climates

How do we know about past climates? Geologists use proxy indicators to understand past climate. A **proxy indicator** is a biological, chemical, or physical signature preserved in the rock, sediment, or ice record that acts like a “fingerprint” of something in the past [42]. Thus they are an indirect indicator of something like climate. For ancient glaciations from the Proterozoic and Paleozoic, there are rock formations of glacial sediments such as the diamictite (or tillite) of the Mineral Fork Formation in Utah. This dark rock has many fine-grained components plus some large out-sized clasts like a modern glacial till [43; 44].

For climate changes during the Cenozoic Era (the last 65 Ma), there is a detailed chemical record from the coring of deep-sea sediments as part of the Ocean Drilling Program. Studies of deep-sea sediment use stable carbon and oxygen isotopes obtained from the shells of deep-sea benthic foraminifera that have settled on the ocean floor over millions of years. Oxygen isotopes are a proxy indicator of deep-sea temperatures and continental ice volume [45].
Sediment Cores – Stable Oxygen Isotope

Oxygen isotopes are an indicator of past climate. The two main stable oxygen isotopes are $^{16}$O and $^{18}$O. They both occur in water ($\text{H}_2\text{O}$) and in the calcium carbonate (CaCO$_3$) shells of foraminifera as the oxygen component of both of those molecules. The most abundant and lighter isotope is $^{16}$O. Since it is lighter, it evaporates more easily from the ocean’s surface as water vapor, which later turns to clouds and precipitation on the ocean and land.

During geologic times when the climate is cooler, more of this precipitation is locked onto land in the form of glacial ice. Consider the giant ice sheets, more than a mile thick, that covered a large part of North America during the last ice age only 14,000 years ago. During glaciation, the glaciers effectively lock away more $^{16}$O, thus the ocean water and foraminifera shells become enriched in $^{18}$O. Therefore, a ratio of $^{18}$O to $^{16}$O in calcium carbonate shells of foraminifera is an indicator of
past climate. The sediment cores from the Ocean Drilling Program record a continuous accumulation of sediment.

**Sediment Cores – Boron Isotopes and Acidity**

Boron-isotope ratios in ancient planktonic foraminifera shells in deep-sea sediment cores have been used to estimate the pH (acidity) of the ocean over the past 60 million years. Ocean acidity is a proxy for past atmospheric CO$_2$ concentrations. In the early Cenozoic, around 60 million years ago, CO$_2$ concentrations were over 2,000 ppm and started falling around 55 to 40 million years ago possibly due to reduced CO$_2$ outgassing from ocean ridges, volcanoes, and metamorphic belts and increased carbon burial due to uplift of the Himalaya Mountains. By the Miocene (about 24 million years ago), CO$_2$ levels were below 500 ppm [46] and by 800,000 years ago CO$_2$ levels didn’t exceed 300 ppm [47].

**Carbon Dioxide Concentrations in Ice Cores**

For the more recent Pleistocene climate, there is a more detailed and direct chemical record from coring into the Antarctic and Greenland ice sheets. Snow accumulates on these ice sheets and creates yearly layers. Ice cores have been extracted from ice sheets covering the last 800,000 years. Oxygen isotopes are collected from these annual layers and the ratio of $^{18}$O to $^{16}$O is used to determine temperature as discussed above. In addition, the ice traps small atmospheric gas bubbles as the snow turns to ice.
Small pieces of this ice are crushed and the ancient air extracted into a mass spectrometer that can detect the chemistry of the ancient atmosphere. Carbon dioxide levels are recreated from these measurements. Over the last 800,000 years, the maximum carbon dioxide concentration during warm times was about 300 ppm and the minimum during cold stretches was about 170 ppm [46; 47; 48]. The carbon dioxide content of earth’s atmosphere is currently over 400 ppm.

![Composite CO2 record (0-800 kyr BP)](image)

**Figure 1**: Composite carbon dioxide record from the last 800,000 years based on ice core data from EPICA Dome C Ice Core.

**Oceanic Microfossils**

Microfossils, like foraminifera, diatoms, and radiolarians, can be used to interpret past climate records. In sediment cores, different species of microfossils are found in different layers. Groups of these microfossils are called assemblages. One assemblage consists of species that lived in cooler ocean water (in glacial times) and another assemblage found at a different level in the same sediment core is made of warmer water species [49].
Tree Rings

Figure 1: Tree rings form every year. Rings that are farther apart are from wetter years and rings that are closer together are from dryer years.

Every year a tree will grow one ring with a light section and dark section. The rings vary in width. Since trees need a lot of water to survive, narrower rings indicate colder and drier climates. Since some trees can be several thousands years old, we can use their rings for regional paleoclimatic reconstructions. Further, dead trees such as those used in Puebloan ruins can be used to extend this proxy indicator, which showed long term droughts in the region and why their villages were abandoned.

Figure 1: Summer temperature anomalies for the past 7000 years (Source: R.M.Hantemirov)

Pollen

Figure 1: Scanning electron microscope image of modern pollen with false color added to distinguish plant species. (Source: Dartmouth Electron Microscope Facility, Dartmouth College)

Flowering plants produce pollen grains. Pollen is distinctive when viewed under a microscope. Sometimes pollen can be
preserved in lake sediments that accumulate every year. Coring of lake sediments can reveal ancient pollen. Fossil pollen assemblages are groups of pollen from multiple species such as spruce, pine, and oak. Through time (via the sediment cores and radiometric age-dating techniques), the pollen assemblage will change revealing the plants that lived in the area at the time. Thus the pollen assemblages are an indicator of past climate since different plants will prefer different climates [50]. For example, in the Pacific Northwest east of the Cascades, a region close to the border of grasslands and forest, a study tracked pollen over the last 125,000 years covering the last two glaciations. As shown in the figure (Fig. 2 from reference Whitlock and Bartlein 1997 [51]), pollen assemblages with more pine tree pollen are found during glaciations and pollen assemblages with less pine tree pollen are found during interglacial times [51].

Other Proxy Indicators

Paleoclimatologists study many other phenomena to understand past climates such as human historical accounts, human instrument record from the recent past, lake sediments, cave deposits, and corals.

References


