Chapter 4: Climate Change Through Earth History

1. Why Past Climate Change Matters

A frequent question to geologists who talk about climate change is why current climate change matters if change has been occurring throughout Earth's history. The practical answer to that question is that current change is rapid and significant enough to matter to *people* and it matters to other living things, which are also valued by people. But why should we care about climate change that has occurred in the Earth's history?

One answer is pedagogical: Though understanding ancient climates is not likely to be the most important thing for students to know about current climate change, it may help students see that climates *can* change, put the kinds of changes we see today into a historical perspective, and help students understand how researchers use paleoclimates to study our currently changing climate. The idea that the Earth's climate could potentially change is an abstract concept that is outside the range of our personal experiences and was, until a couple hundred years ago, a radical idea. Accepting the idea that the Earth does change has profound implications for how we see the Earth and its future. Seeing direct evidence of Earth change in one's own region—such as through rocks or fossils normally associated with much colder or warmer, or drier or wetter, environments than occur now—communicates the idea that the Earth *does* change. It also connects climate to aspects of Earth systems such as the rock and fossil record.

Another answer is scientific and practical: Past climates help scientists understand how the Earth *could* change by understanding how the Earth *has* changed. Climate scientists' predictions for temperature and precipitation changes associated with current climate change frequently rely on sophisticated computer models. But there is no practical way to physically test hypotheses derived from such models about the long-term rates of glacial retreat, changes in oceanic circulation, influences on organisms, and so on—we can't recreate a global laboratory except in a computer simulation. Climate change events in Earth's history, however, have performed some of the experiments for us. The lessons learned from ancient climates may be difficult to apply to modern climate change because the circumstances (land positions, atmospheric chemistry, vegetation, and so on) become increasingly different the further back we go in time, but even very ancient climate changes in a world that seems quite foreign provide sometimes surprising lessons about how the Earth system operates, and how fast and to what extremes it can change.

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A Very Short Guide to Climate Change

Climate and Earth History

phytoplankton • *one-celled photosynthetic algae that float near the surface of bodies of both marine and freshwater.*

Paleozoic • *a geologic time era that extends from 541 to 252 million years ago. Fossil evidence shows that during this time period, life evolved in the oceans and gradually colonized the land.*

In simplest terms, detecting climate change of the past requires only a sequence of sediments. For those of us with an outcrop of sedimentary rocks in our area, the results of climate change are within view nearly any time we see two or more layers that look different. Little specialized knowledge is necessary: the fact that the rocks vary in color, resistance to weathering, or bedding patterns indicates that the character of the sediments changed, and that means we are seeing the results of past environmental change. There is a good chance that this environmental change was associated with, if not caused by, changes in climate. Add in understanding of a few principles about how sediments vary among environments, and you and your students can hypothesize about climate changes in the geologic past wherever you may find sedimentary rocks.

It's also interesting to ponder that the rocks we see that are a record of climate change frequently played a role in creating that change: the carbon stored in sediments and rocks is part of the global carbon cycle that affects atmospheric CO_2 concentrations. Large amounts of Earth's carbon are tied up in limestones ($CaCO_3$) and organic carbon in sedimentary rocks, particularly shales and coal, which you may be able to see at the surface in your region. In many areas of the country we can also observe such deposits in the making: organic-rich sediment in modern environments around us, e.g., dark muds along ponds and lakes, and accumulation of peat in swampy areas. These are effectively the same substances that became, after the pressure and heat of deep burial, fossil fuels. Mass production of energy through the burning of fossil fuels is burning the accumulation of hundreds of thousands or millions of years of forests and **phytoplankton**. It may be easier for students to put into perspective how rapid must be the rate of change of CO_2 in the atmosphere today we when think about how long it took for the accumulation of organic carbon that became fossil fuels.

2. Observing Climate Through Time in the Rock Record

When we use the term "climate change" we are referring to a global average—"global warming"—while also referring to other environmental shifts in specific geographic regions. Though we hear particularly about warming, we know that for some places the most significant trends impacting living things may be changes in storm intensity or precipitation in addition to or instead of temperature changes. Some places in the world may experience cooling (at least for some years) even while most places are experiencing warming. The same is true of trends in climate history: the rock records in different parts of the United States reflect environmental changes in those regions, which may or may not clearly reflect global changes occurring at the time the rocks were deposited.

Another consideration is that the history of climate at any specific place over geologic intervals of hundreds of millions of years will involve changes in latitudinal position of the continent. To make sense of why a place had the climate it did at some time in the past we must distinguish between the roles of moving tectonic plates from the role of changing global climates. Rocks and fossils from the **Paleozoic** era (*Figure 4.1*) in the United States primarily indicate warm environments, but this can be explained from independent evidence that North America was at low latitudes, even right over the equator, during

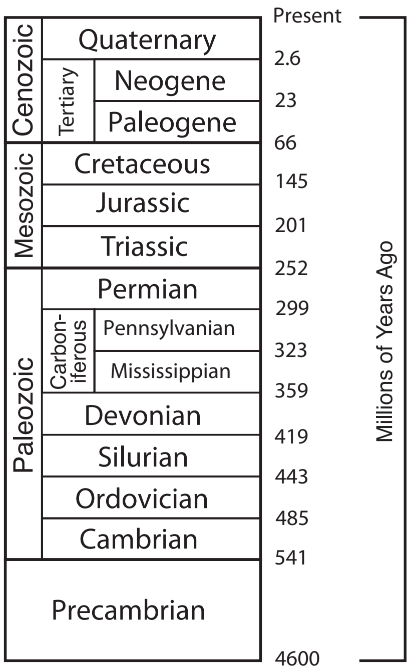


Figure 4.1: The geologic time scale (spacing of units not to scale).

About the time scale

The time scale in The Teacher-Friendly Guides™ follows that of the International Commission on Stratigraphy (ICS). The Tertiary period, though it was officially phased out in 2008 by the ICS, remains on the scale in the Guides, since "Tertiary" is found extensively in past literature. In contrast, the Carboniferous and Pennsylvanian & Mississippian periods all enjoy official status, with the latter pair being more commonly used in the US.

much of that time. We use evidence from other continents that were at higher latitudes during the Paleozoic to ascertain that some geologic time intervals were relatively warm at a global scale, but other time intervals were relatively cool with polar glaciers (such as the end-Ordovician and the Carboniferous periods).

These complications notwithstanding, the ancient environments we can see in the rock record of the United States, in particular those in one's own region, may provide useful discussion points about how the Earth's climate changes and what those changes mean for current climate change.

2.1. Inferring Ancient Climates

How do we know what ancient climates were like? To know the average temperature of the world 10,000 years ago, since we cannot look at a thermometer, we need a substitute—a **proxy**—that indirectly recorded that information.

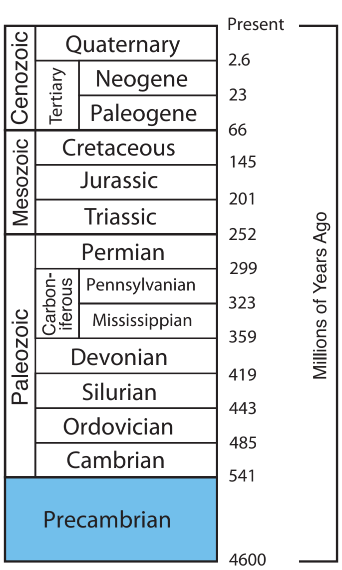
proxy - an alternative to direct measurements of climate variables; data from sources like tree rings, lichens, and pollen are used to infer climate information.

Wherever Earth's atmosphere contacts water and sediment (stirring it, heating and cooling it) and helps or hinders the growth of organisms, climate records are left behind. Earth scientists reconstruct ancient climates by using traces left in the rocks, fossils, and sediments available on the Earth's surface. Even after thousands or millions of years, many of these materials contain information about the environmental conditions that existed at the time that they were laid down as soils or preserved in sediments in bodies of water. This climatic information can be found in unconsolidated sediments (for example, in mud at the bottom of a pond), in rocks and fossils, in glacial ice sheets, or even in a living tree or coral colony. Each of these systems records something about the world in which they formed. See *Boxes 4.1* (proxies from rocks), *4.3* (fossils), *4.4* (oxygen isotopes), *4.7* (ice cores), and *4.8* (living organisms) for more detailed information.

Precambrian ◦ a geologic time interval that spans from the formation of Earth (4.6 billion years ago) to the beginning of the CAMBRIAN (541 million years ago). Relatively little is known about this time period since very few fossils or unaltered rocks have survived. What few clues exist indicate that life first appeared on the planet as long as 3.9 billion years ago in the form of single-celled organisms.

bacteria ◦ single-celled microorganisms with cell membranes but without organelles or a nucleus.

protists ◦ a diverse group of single-celled eukaryotes (organisms with complex cells containing a nucleus and organelles).



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Of course, not all of these proxy materials are present everywhere, and in fact many places have few or none. Not unlike human history, we have to piece together geological history from different times and different places, to make a general storyline across broad regions or globally.

3. History of the Earth's Climate

Though a minor amount of Earth's internal heat is released at the surface, nearly all heat that influences climate comes from the sun. The heat of the sun on the Earth's surface varies through geologic time because of variations in the sun itself and predictable variations in the way the Earth tilts, rotates, and revolves around the sun. Just as important as heat received at the Earth's surface, however, is the relative proportion in which that heat is reflected away, absorbed but quickly lost back to space, or "trapped" by the atmosphere. Broadly speaking, factors affecting retaining or losing heat from the sun has been the driver of most global climate change over the history of the Earth.

Changes in concentrations of the greenhouse gases CO_2 and CH_4 had an impact on Earth's climate from early in Earth history, and reciprocal variations in CO_2 and O_2 have characterized some of the largest events in the history of both Earth and life. Changes to the Earth's surface have also had a big effect on the amount of heat the Earth and ultimately the atmosphere absorb. Surface phenomena such as moving continents, changing rocks at the surface, evolution of plant life onto land, and changes in distribution of ice explain why Earth's climate has changed over geologic time in the way that it has. The insights this gives climate scientists regarding **forcings**, **feedbacks**, interactions, and sensitivities of the climate system can be applied to understanding current and future human-induced climate changes.

3.1. The Early Evolution of the Atmosphere

The first four billion years (about 85%) of Earth history, collectively known as the **Precambrian**, might be described as the interval during which Earth systems came to be (relatively speaking) the way they are today: for example, plate tectonics, atmospheric chemistry and structure, and ecosystem processes developed over the course of that interval. The Precambrian became, sometime within its first third, occupied by a great diversity of **bacteria**, with **protists** diversifying in the final third.

The early evolution of Earth's atmospheres signals the major influence CO_2 and CH_4 would go on to play in the evolution of climate and life through Earth history. Most people may not stop to consider whether an atmosphere like the one we have is inevitable. Every planet in our solar system, or any planetary system for that matter, will have its own unique chemical composition. If circumstances are such that a planet retains an atmosphere, it will likely have started with gasses made of some combination of the elements hydrogen, helium, carbon, nitrogen, and oxygen. Smaller planets, like Mars, as well as moons in our own solar system, have a very thin atmosphere: at one time there were greater quantities of gases on these planetary bodies, but the gravity of the planet did not retain the same mass of air as the Earth has.

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Not long after the Earth first formed, more than 4.5 billion years ago, its atmosphere was composed mostly of hydrogen and helium, which, because of the Earth's modest size and gravity field, was mostly lost to space. Volcanic activity ("degassing") and to a much lesser extent, collisions with meteorites and comets added water vapor, carbon dioxide, and nitrogen to the atmosphere. As the Earth cooled enough for liquid water to form, the vapor formed clouds from which torrential rains poured for millions of years, absorbing salt and other minerals from the earth as the rainwater coursed to the lowest areas, forming Earth's oceans and seas.

The Earth still could have lost its atmosphere, in spite of its gravity: **ionizing radiation**—"solar wind"—from the sun might have, over time, knocked most gas molecules out of the atmosphere. The Earth, however, has a **magnetic field** associated with convection in its core and this magnetic field, originating sometime in Earth's first billion years, has since acted to block most of the effects of ionizing radiation.

Within the first billion years of Earth's history, as the early atmosphere was evolving, the surface of the Earth was cooling to form a solid crust of rock (there are mineral crystals indicating that this process may have started as early as 4.4 billion years ago). Regardless of precisely when this took place, it represented the formation of continental terranes that were the precursor to the processes of plate tectonics that have continued ever since. The motion of these plates through different latitudinal climate zones, and the size and arrangement of the continents, have greatly impacted heat retention and patterns of circulation and precipitation. The amount and types of minerals at the Earth's surface exposed to the atmosphere (or covered by water or glaciers) played a huge role in atmospheric chemistry. For example, rock that is enriched in **organic matter** will release abundant amounts of carbon dioxide as it weathers, while rock rich in **feldspar** and mica will remove carbon dioxide during the chemical process of weathering.

UV light caused much of the atmospheric gas in the form of methane (CH_4) and ammonia (NH_3) to dissociate, leaving N_2, CO_2, and H_2, the latter of which was lost to space. Molecular oxygen (O_2) did not exist, and thus was not available to oxidize surviving molecules such as CH_4. It is widely accepted that energy from the sun very early in Earth history was about 30% less than it is today and, all else being equal, one would expect Earth to have been about 30 °C colder, which should have led to widespread and long-term ice formation. Geological evidence (such as record of liquid water) suggests, however, that this was not the case, thus it seems likely that greenhouse gases—carbon dioxide and the remaining methane, possibly at concentrations that were orders of magnitude greater than in the atmosphere today—acted to maintain a relatively warm Earth.

Over Earth's history, the atmospheric content of N_2 has steadily increased through volcanic degassing. Today it represents over 3/4 of the Earth's atmosphere by volume. N_2 is very stable and as such does not react much with either the rocky surface of the Earth or other molecules in the atmosphere, which allows it to accumulate over time.

Today, by far, the 2nd most abundant gas is O_2, over ¼ of the atmosphere by volume, but it wasn't always this way. The Earth had very little free oxygen until

Precambrian

ionizing radiation • *high-energy electromagnetic energy which can cause ionization in the material through which it passes, for example, x-rays and gamma rays.*

magnetic field • *a conceptualization of the strength and direction of the magnetic force at a distance from an object.*

organic matter • *decomposed remains of plants, animals, and their wastes.*

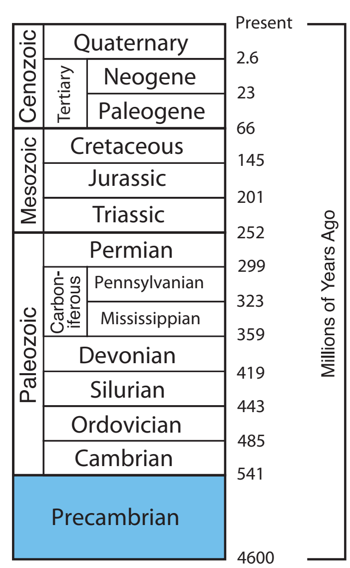

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Figure 4.2: Banded iron formation (Fortescue Falls, Karijini National Park in Western Australia).

the evolution of photosynthesis in bacteria, perhaps beginning about 3.5 billion years ago. This would be one of the first of many instances of life changing the atmosphere. The abundant iron and organic matter in the environment quickly reacted with the oxygen they produced, but after hundreds of millions of years, these oxygen-absorbing sinks (such as extensive deposits of **iron oxide minerals** deposited in **"banded iron" formations**; *Figure 4.2 and Box 4.1*) were exhausted, and free oxygen built up in the atmosphere.

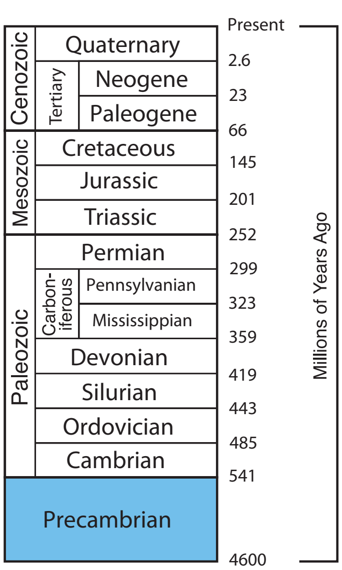
feldspar - an extremely common group of rock-forming minerals found in igneous, metamorphic, and sedimentary rocks. There are two groups of feldspar: alkali feldspar (which ranges from potassium-rich to sodium-rich) and plagioclase feldspar (which ranges from sodium-rich to calcium-rich). Potassium feldspars of the alkali group are commonly seen as pink crystals in igneous and metamorphic rocks, or pink grains in sedimentary rocks. Plagioclase feldspars are more abundant than the alkali feldspars, ranging in color from light to dark. Feldspars are commercially used in ceramics and scouring powders.

iron oxide minerals - a range of minerals containing chemical compounds of iron and oxygen.

The increase in oxygen allowed the development of **ozone** in the **stratosphere**. The ozone layer blocks **ultraviolet light**, and its development may have decreased cell damage in microbial life near the surface. Stratospheric ozone also has a fundamental impact on the structure of the atmosphere. Ozone is responsible for the increase in temperature in the stratosphere with altitude because it absorbs the short wave radiation from the sun; this contributes the relative stability of the stratosphere, above the complex convection and weather of the **troposphere**.

The timing of extensive iron oxide deposition occurred about the same time as development of extensive glaciation 2.4 to 2.2 billion years ago, and it has been suggested that increased oxygen reacted with the greenhouse gas methane, converting it into carbon dioxide, a less effective greenhouse gas. This cooling is evidenced by globally distributed glacial deposits, some of which are thought to have occupied low (equatorial) latitudes. This glacial interval is known as the Huronian (named after deposits in Michigan). A significant fraction of the Earth's land may have been covered in ice for as long as 300 million years. At that time the continental plates made up less than half as much of the Earth's surface as they do today and were unified as a continent known as Arctica.

An ice-covered planet would remain that way because almost all of the sun's energy would be reflected by the ice back into space, but this did not happen on Earth, probably because of plate tectonics. The glaciation was eventually disrupted by ongoing volcanic activity, which added carbon dioxide and methane back into the atmosphere. These gases are usually removed from the atmosphere by organisms and the weathering of rocks, but these processes would have stopped while the continents and oceans were covered in glacial ice. After millions of years, the concentrations of methane and carbon dioxide increased to the point that greenhouse warming began to melt the ice



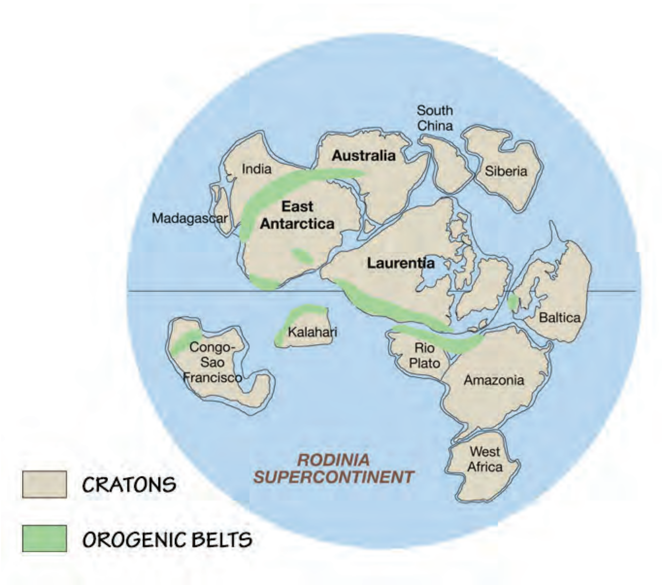


Figure 4.3: The supercontinent Rodinia, circa 1.1 billion years ago. Laurentia represents proto-North America. (See Teacher-Friendly Guide website for a full color version.)

sheets. Once the melting started, more of the sun's energy was absorbed by the surface, and warming feedbacks began. Because the oceans had been covered, nutrients (like the mineral phosphorous) from chemical weathering of the rocks accumulated in the oceans. Population explosions of cyanobacteria used these nutrients to produce more and more oxygen capable of combining with freshly thawed carbon sources to make more carbon dioxide, further enhancing the warming. The oxygen release became part of a relatively rapid increase in atmospheric O_2.

Another very extensive glaciation occurred in the late Precambrian, about 717 million years ago, during the Cryogenian.[1] There is evidence suggesting that the entire surface of the planet became covered in ice, a hypothesis called "Snowball Earth," possibility involving cycles of decline and increase in greenhouse gases similar to those hypothesized for the Huronian glaciation. The North American portion of the supercontinent **Rodinia**, which had formed by 1.1 billion years ago, was near the equator and in the center of the supercontinent (Figure 4.3). Two extensive phases of glaciation occurred during this time, called the **Sturtian glaciation** (about 717 to 660 million years ago) and, as Rodinia began to break up, the **Marinoan glaciation** (about 640 to 635 million years ago) (*Figure 4.4*). The fact that North America was at such a low latitude, yet had glaciers (based, for example, on deposits in Idaho and Utah), is strong evidence that the Earth was cold enough to have experienced ice at a global scale.[2]

[1] There remains uncertainty about the degree to which the Earth was frozen over during the interval of "Snowball Earth," but it's clear that glaciation extended to the equator.

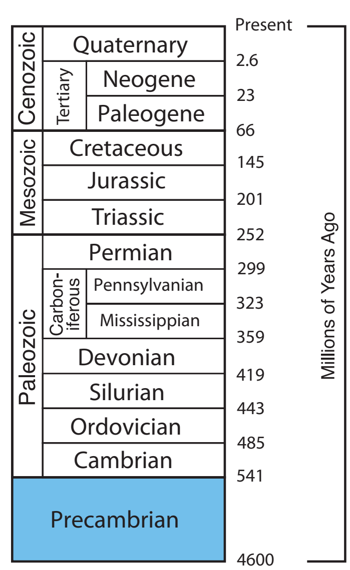
[2] The term "ice age" has different connotations depending on context. A common usage of the term refers to the whole time interval of large scale glacial-interglacial cycles in the Pleistocene (2.6 million years). The public usually thinks specifically of the most common glacial advance (the Last Glacial Maximum) about 20,000 years ago among a series of glacial-interglacial cycles. A broader definition is an interval over which there exists substantial glacial ice at the poles (or beyond).

"banded iron" formation ◦ geologic formations comprised of layers of PRECAMBRIAN, iron-rich, sedimentary rock.

ozone ◦ a molecule (O_3) found in the STRATOSPHERE which absorbs ultraviolet light. When found near the surface of the Earth, ozone is considered a pollutant because it is a component of smog and can cause lung irritation.

stratosphere ◦ the second layer above the Earth's surface in the ATMOSPHERE. The stratosphere reaches to about 50 kilometers (30 miles) above the Earth's surface.

ultraviolet light ◦ electromagnetic radiation in the part of the spectrum with wavelengths from 10 to 400 nanometers.



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troposphere • *the layer of the ATMOSPHERE extending from the Earth's surface to about 7 to 20 kilometers (4 to 12 miles) above the surface. The height of the troposphere depends upon latitude and season.*

Rodinia • *a supercontinent that contained most or all of Earth's landmass, between 1.1 billion and 750 million years ago, during the PRECAMBRIAN. Geologists are not sure of the exact size and shape of Rodinia. It was analagous to but not the same supercontinent as PANGAEA, which formed several hundred million years later during the PERMIAN.*

Sturtian glaciation • *a time in Earth's history, around 717 to 660 million years ago, when the entire planet may have been covered in ice.*

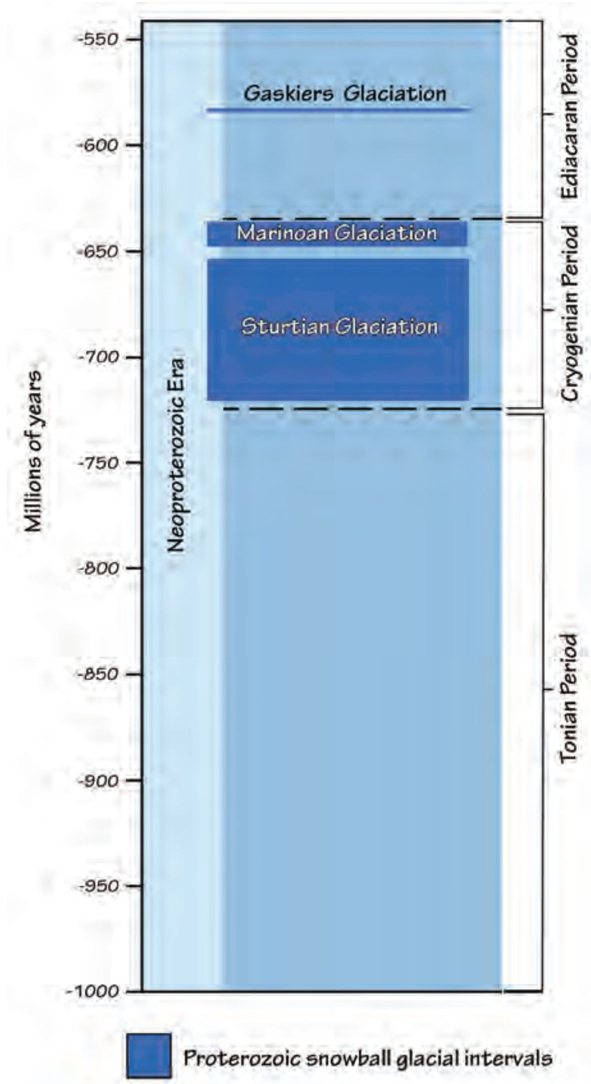


Figure 4.4: Snowball Earth periods during the late Precambrian.

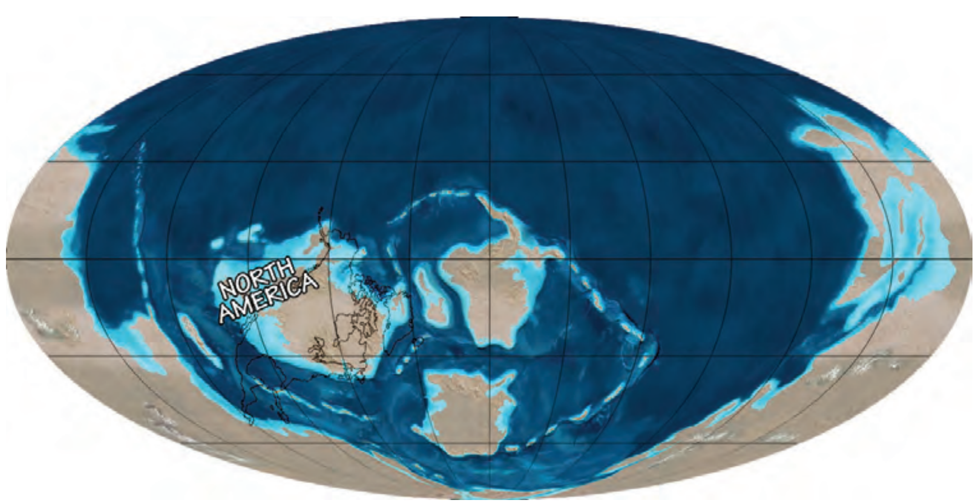
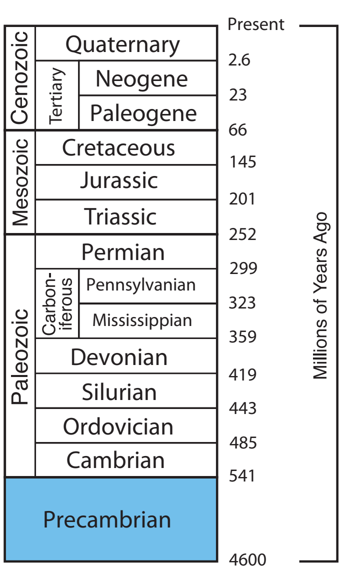


Figure 4.5: Earth during the early Cambrian, around 545 million years ago.



By 635 million years ago, the Earth had warmed again, and the North American continent had moved towards the equator (*Figure 4.5*). About this time we find some of the first animal fossils. It isn't clear what is the causal relationship between these major climate perturbations and major events in evolution.

3.2. The Early to Mid Paleozoic Era

The interval that covers, except for the the very beginning, most of the history of animal life, is known as the **Phanerozoic eon**; its history, particularly with respect to climate, is far better known scientifically than the Precambrian and far more commonly covered in science education and popular media. The Phanerozoic is split into the Paleozoic, Mesozoic, and Cenozoic eras.

From the **Cambrian** to **Silurian** animals primarily diversified in the seas; the Devonian period was an important transition, as the first sizable (but still coastal) forests and land vertebrate communities evolved. The early Paleozoic era is interesting from a climate perspective in part because there are numerous repeated patterns of changing climate and sea level associated with increased or decreased rates of evolution and extinction. Across the US there are opportunities to observe how communities of organisms responded (for example, in species composition, abundance, and size) from layer to layer, in fossil-rich marine sedimentary rocks. These patterns may give us clues to how marine organisms will respond long-term to current climate change.

3.2.1 Cambrian and Ordovician Periods

With the start of the Paleozoic era, global climates across the world were warm, and North America was located in the low, warm latitudes of the Southern Hemisphere. What would become the northern US was located just north of the equator. Broadly speaking, we find sedimentary deposits in North America throughout the Paleozoic Era that reflect tropical conditions (see *Box 4.1*). These deposits say more about the position of North America near the equator than they do about global climates, which varied widely through the Paleozoic. Evidence for warm climates in the Cambrian and **Ordovician** periods we see today in extensive limestone deposits and ancient reefs (see *Box 4.3*), for example, on the western (California and Nevada) and eastern (New York and Pennsylvania) side of the continent and in the Midwest. For some time in the Ordovician much of the Midwest was covered by very pure, quartz-rich sand, which suggests that the climate was intensely wet and warm and that the sand was washed or blown (or both) back and forth for a long time before being buried.

The Earth went through another ice age from 460 to 430 million years ago (*Figure 4.6*). The continent of **Gondwana** (modern South America, Africa, Australia, Antarctica, Arabia, and India) was located over the South Pole and became covered in glacial ice. This led to global cooling, which was associated with the first of five major **mass extinctions** that have occurred over the last half-billion years.

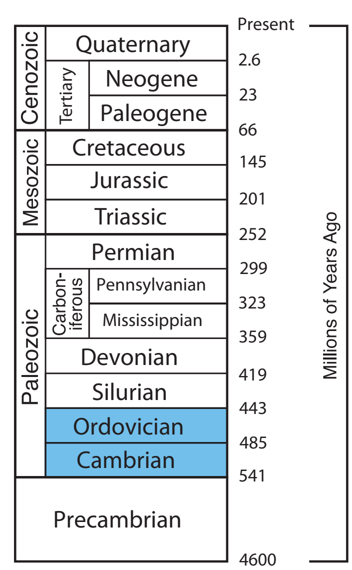
During the Phanerozoic not only has polar glaciation not extended to equatorial regions, but equatorial regions remained warm. For example, during this glacial interval, low latitude reefs grew around the shallow edges of a wide basin

Marinoan glaciation • a time in Earth's history, around 640 to 635 million years ago, when the entire planet may have been covered in ice.

Phanerozoic eon • a geological eon representing the entirety of geological history after the PRECAMBRIAN, from 541 million years ago to the present.

Cambrian • a geologic time period lasting from 541 to 485 million years ago. During the Cambrian, multicellular marine organisms became increasingly diverse, as did their mineralized fossils. The Cambrian is part of the PALEOZOIC Era.

Silurian • a geologic time period spanning from 443 to 419 million years ago. During the Silurian, jawed and bony fish diversified, and life first began to appear on land.

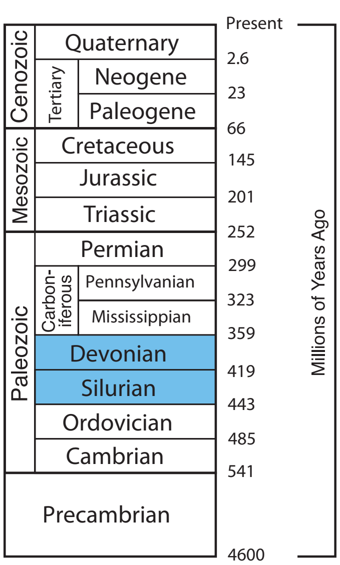


Early to Mid Paleozoic

Ordovician - a geologic time period spanning from 485 to 443 million years ago. During the Ordovician, invertebrates dominated the oceans and fish began to diversify.

Gondwana - the supercontinent of the Southern Hemisphere, composed of Africa, Australia, India, and South America. It combined with the North American continent to form PANGAEA during the late PALEOZOIC.

mass extinction - the extinction (loss of the last living member of a species) of a large percentage of the Earth's species over a relatively short span of geologic time.



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Box 4.1: Proxies from rocks

Sedimentary rocks are formed through breakdown of other rocks into sediment, which is then transported and deposited by wind or water. When the sediment is compressed or cemented and turned into rock, it retains clues about the environment in which it formed. By observing modern oceans, for example, scientists note that limy sedi¬ments and reefs (composed of calcium carbonate) usually accumulate in warm, shallow, clear seawater, and they then use this to conclude that ancient carbonates might have formed in similar environments.

Chemical elements in rocks, and even in some fossils, can also record information about the environment at the time that the rocks were formed. Particularly useful for recreating ancient climates are the different forms (or **isotopes**) of the element oxygen (see *Box 4.4* on isotope proxies).

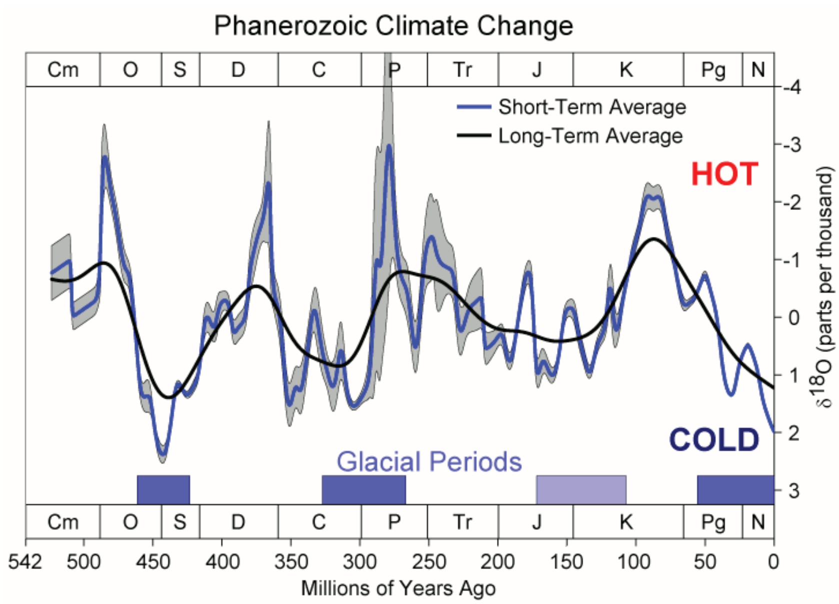

Figure 4.6: Changing global climate throughout the last 542 million years. These data were compiled using the ratios of stable oxygen isotopes found in ice cores and the carbonate skeletons of fossil organisms. (See Teacher-Friendly Guide website for a full color version.)

centered in Michigan. These reefs were among the largest the world had ever seen and today the remains of the reefs (as limestone) can be found across much of the Midwest with the thickest deposits occurring in Indiana and Illinois.

3.2.2 Silurian and Devonian Periods

From 430 to 300 million years ago, North America moved north across the equator (*Figure 4.7*), and the cycle of warming and cooling was repeated again. Silurian deposits of salt in Michigan and New York indicate that the North American climate experienced dry climates and restricted circulation during a warm interval beginning 430 million years ago. Eventually, the salinity in the shallow seas of the ancient Midwest and Northeast returned to normal in the **Devonian**, when sea level rose. A diverse warm water reef fauna occupied the sea floor of shallow seas over broad swaths of the East and Midwest.

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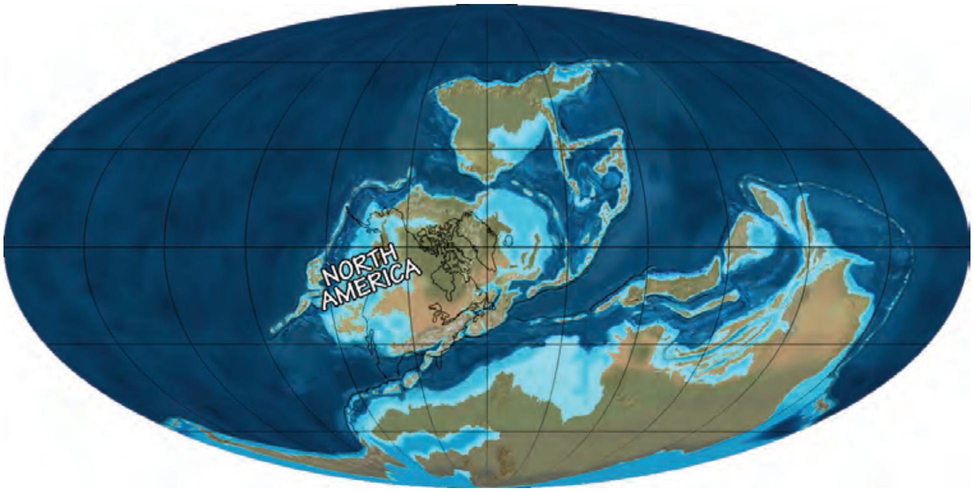


Figure 4.7: Earth during the late Devonian, about 370 million years ago.

In the Devonian a variety of tectonic changes occurred that led to the formation of continental basins with plankton productivity so high that their decay led to depletion of oxygen from the seafloor and sediments. The lack of oxygen allowed organic matter to accumulate instead of decaying, leading to the deposition of black, carbon-rich shale (see *Box 4.2*). Though all geologic periods have experienced such deposits, some Devonian-age marine rocks that are currently especially important sources for natural gas and petroleum include the Barnett shale (primarily northern Texas), Marcellus shale (especially Pennsylvania), and the Bakken Formation (especially North Dakota).

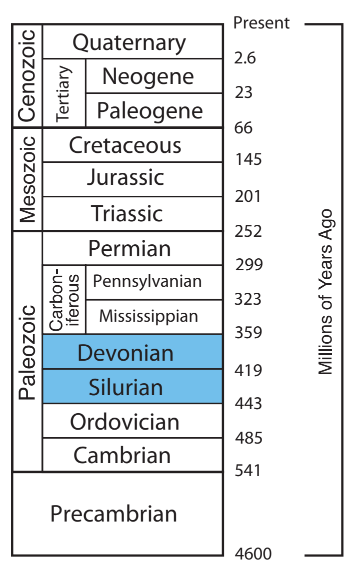
At the end of the Devonian the fauna suffered a mass extinction that eliminated many of the more important groups of reef-builders and other animals that occupied the shallow seas. The causes of this mass extinction, which actually occurred in a series of steps, are still uncertain. Dropping sea levels and cooling climates as the Earth entered another glacial interval have been implicated.

Box 4.2: Earth system links between ancient and modern climates

Throughout the Phanerozoic eon there have been circumstances during which large amounts of marine organic matter, particularly phytoplankton, were deposited in sediment before decaying. The decay rate of organic matter is controlled by the amount of oxygen in the bottom water and surface sediment, which itself is controlled by bottom circulation and quantity of organic matter decaying. When the amount is great, oxygen-loving bacteria use up the available O_2 faster than it's replenished. The process is affected by climate indirectly in the sense that temperature, nutrient availability and light are influenced by climate phenomena. In turn, the buried organic matter derived from photosynthesis removes carbon from the atmosphere and thus atmospheric CO_2. Such organic carbon, after being subjected to heat and pressure under additional sediments, became the source for petroleum and natural gas. Burning these fossil fuels releases the "fossil sunshine" and CO_2 that had been stored in rocks beneath the surface.

isotope • *a form of a chemical element that contains a specific number of neutrons. For example, the isotope of carbon with six neutrons is known as carbon-12 (^{12}C) and the isotope of carbon with eight neutrons is carbon-14 (^{14}C). All isotopes of an element contain the same number of protons.*

Devonian • *a geologic time period spanning from 419 to 359 million years ago. The Devonian is also called the "age of fishes" due to the diversity of fish that radiated during this time. On land, seed-bearing plants appeared and terrestrial arthropods became established. The Devonian is part of the Paleozoic.*



Climate and Earth History

3.3 Late Paleozoic Era

The late Paleozoic is the interval during which widespread forests colonized the land and, with them, animal faunas, including particularly **arthropods**, but also **vertebrates**. The Paleozoic ended with the great mass extinction in geologic history. The interval is interesting from a climate systems perspective because of the clear influence of CO_2 in influencing the presence or absence of polar glaciation and, at sufficiently high levels of CO_2, in an extinction.

arthropod • an invertebrate animal, belonding to the Phylum Arthropoda, and possessing an external skeleton (exoskeleton), body segments, and jointed appendages. Arthropods include crustaceans, arachnids, and insects, and there are over a million described arthropod species living today. Trilobites are a major group of extinct arthropods.

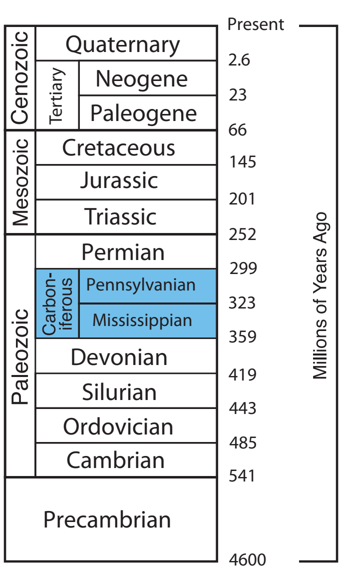
vertebrate • an animal with a backbone, such as fishes, amphibians, reptiles, birds, and mammals.

3.3.1 Carboniferous

The late Devonian and early **Carboniferous** periods were a time of transition for terrestrial ecosystems that had major implications for climate: for the first time, major assemblages of complex land plants, including large plants—trees—developed, first in wet, swampy coastal areas. A combination of the burial conditions and the early, and possibly limited, evolution of organisms that contribute to plant decay led to thick accumulation of organic "**peat**" deposits, trapping organic carbon in what would become coal in places such as southern Illinois, Indiana, Ohio, and western Pennsylvania. The drop in carbon dioxide led to the next glaciation: by the Early Carboniferous, ice capped the supercontinent Gondwana at the South Pole and began to expand northward. Although the Earth's temperature fell during this time and the frozen water trapped in southern hemisphere glaciers caused sea levels to drop, North America remained relatively warm because of its position near the equator. Deposits in the southern part of the Midwest, in particular, show a cyclicity of rising and falling sea level that was caused by advance and retreat of the large ice cap in the Southern Hemisphere.

Carboniferous • a geologic time period that extends from 359 to 299 million years ago. It is divided into two subperiods, the Mississippian and the Pennsylvanian. By the Carboniferous, terrestrial life had become well established.

By the late Carboniferous, North America had collided with Gondwana, eventually leading to the formation of **Pangaea**—a supercontinent composed of nearly all the landmass on Earth (*Figure 4.8*). Pangaea was so large that it created a strong monsoonal climate, much as Asia does today. Large swamps formed along broad floodplains that eventually became the rich coal beds of, for example, Pennsylvania, Tennessee, Kentucky, and West Virginia.



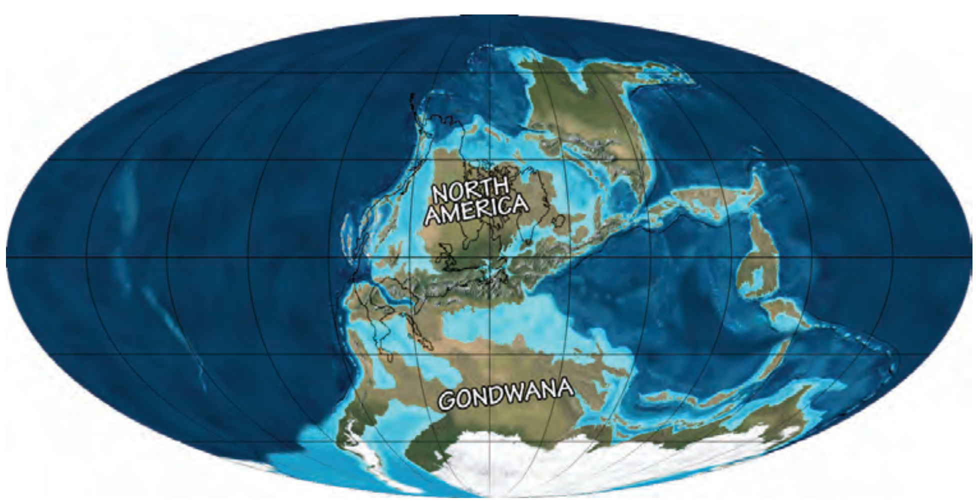


Figure 4.8: Initial formation of Pangaea during the late Carboniferous, around 300 million years ago.

In the late Carboniferous, since the continent was largely tropical, the climate remained warm, despite large southern ice sheets, but the continent had grown much drier. Thick salt deposits accumulated in Utah and Colorado as the seas evaporated. Where the land was exposed, deposits of dust (**loess**) accumulated and were blown across much of the Southwest. The ice age that began in the early Carboniferous lasted well into the **Permian** period, when warm temperatures again became the norm.

3.3.2 Permian

During the Permian, sea level gradually began to decrease, in this case not because of the development of glacial ice, but because of decreases in sea floor spreading associated with the formation of the supercontinent Pangea (*Figure 4.9*). Seafloor spreading (rifting) of hot, mantle-derived rock creates undersea mountain ranges (such as today's **mid-Atlantic Ridge**), which displace ocean water onto the continents. When the plates are connected, as in the supercontinent Pangea, seafloor spreading is reduced, ridges displace less water, and sea level drops.

The climate was drier than that in the Carboniferous, and mudflats with salt and **gypsum** formed across the Southwestern states. Sand dunes started to become widespread (*Figure 4.10*). A shift in plant type—from water-loving ferns and **horsetails** to those better adapted to drier conditions—further suggests a change in climate during the Permian (*Box 4.3*). A large, low-latitude desert formed along Pangaea's western margin, generating extensive dune deposits.

By the end of the Permian, the southern ice sheets had disappeared. As the **Triassic** period began, the Southwestern U.S. moved north from the equator. The world warmed, and would stay warm through the **Mesozoic**. Warm, arid desert conditions existed in the core of the supercontinent, as indicated, for example, by ancient sand dunes preserved in sedimentary rocks.

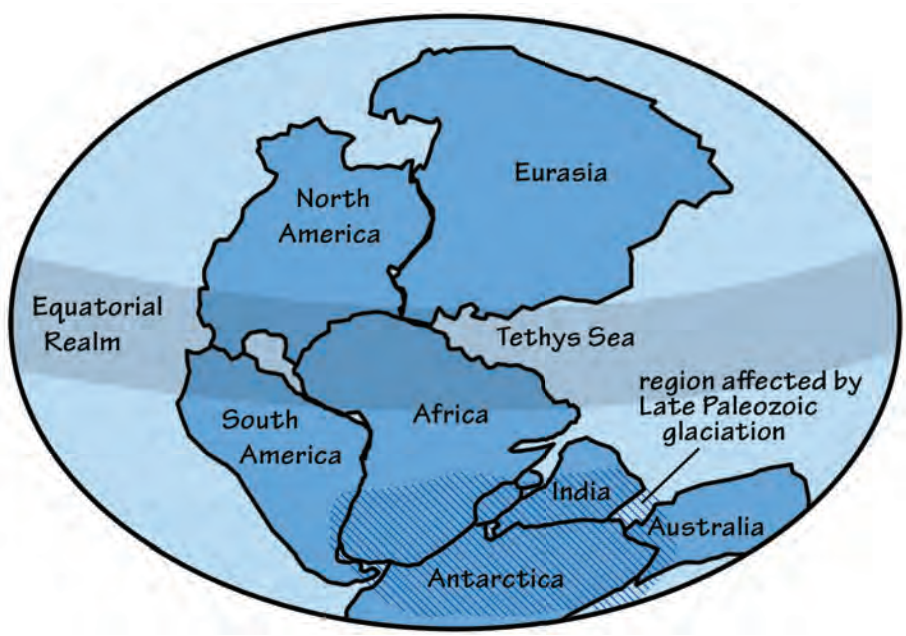
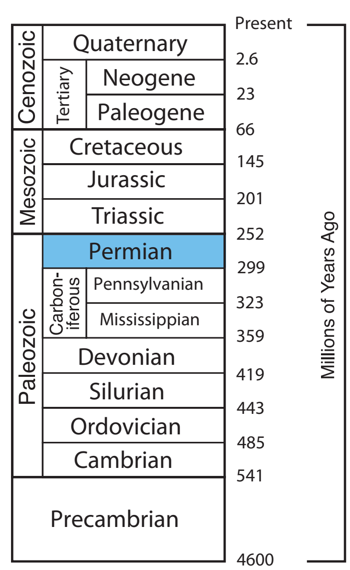


Figure 4.9: Pangaea during the late Paleozoic era.

peat ◦ an accumulation of partially decayed plant matter. Under sufficient heat and pressure, it will turn into lignite COAL over geologic periods of time.

Pangaea ◦ a supercontinent, meaning "all Earth," which formed over 300 million years ago and lasted for almost 150 million years, during which all of the Earth's continents were joined in a giant supercontinent. Pangaea eventually rifted apart and separated into the continents in their current configuration.

loess ◦ very fine-grained, wind-blown sediment, usually rock flour left behind by the grinding action of flowing GLACIERS.



Box 4.3: Proxies from fossils

Fossils—the remains or traces of once-living things preserved in the Earth's crust—can be compared to organisms in modern environments to infer the past environment in which they lived (*Figure A*). For example, fossil fish and seashells can reasonably be assumed to have lived in water, even though the place where the fossils were found is now dry land. Fossil reptiles or palm trees found in what are now much cooler, high-latitude locations testify to these areas once having a much warmer climate. Corals are mostly colonial, marine animals that make hard skeletons out of calcium carbonate ($CaCO_3$). Modern corals live mainly in warm, tropical seas. Fossil corals found today in very different environments, such as upstate New York, are therefore indicative of major changes in the climate of the area.

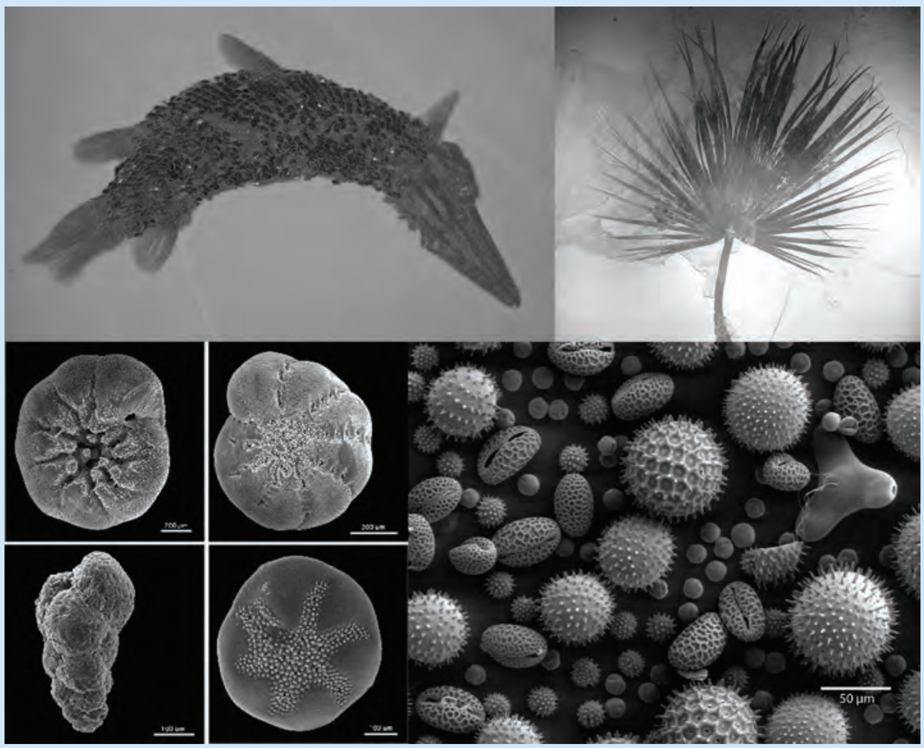


Figure A: Examples of fossil climate proxies. Top left and right are a fossil palm frond and alligator, respectively, both Eocene Epoch, Wyoming. Bottom left shows benthic foraminifera, found in marine sediments; the species are (clockwise from top left) Ammonia beccarii, Elphidium excavatum clavatum, Buccella frigida, *and* Eggerella advena. *Bottom right image shows common pollen grains (greatly magnified), including sunflower and lily pollen.*

Fossil leaves frequently have characteristic shapes that are, in part, the result of the habitat in which they live. Looking at their shape scientifically with a process called **leaf margin analysis** can help to reconstruct ancient environments and climates (*Figure B*). The edges of modern leaves are indicative of their climate and environment; smooth-edged leaves with narrow, pointed "drip tips" at the ends are common in rainforests where they function to rid the leaves of excess water, whereas toothed leaf edges are more common in temperate environments to preserve water. Scientists measure leaves in modern environments and correlate their size, shape,

Permian - *the geologic time period lasting from 299 to 252 million years ago. During the Permian, the world's landmass was combined into the supercontinent PANGAEA. The Permian is the last period of the PALEOZOIC. It ended with the largest mass extinction in Earth's history, which wiped out 70% of terrestrial animal species and 90% of all marine animal species.*

mid-Atlantic Ridge - *a ridge on the floor of the Atlantic Ocean generally running North-South at the boundary of tectonic plates, where these plates are moving apart.*

gypsum - *a soft, sulfate mineral that is widely mined for its use as fertilizer and as a constituent of plaster. Alabaster, a fine-grained light colored variety of gypsum, has been used for sculpture making by many cultures since ancient times.*

horsetail - *a terrestrial plant belonging to the Family Equisetaceae in the plant division Pteridophyta, and characterized by hollow, jointed stems with reduced, unbranched leaves at the nodes.*

Triassic - *a geologic time period that spans from 252 to 201 million years ago. During this period, DINOSAURS, PTEROSAURS, and the first mammals appear and begin to diversify.*

Box 4.3: Continued

and edge appearances with the temperature and humidity of the region. That information can then be applied to fossil leaf measurements in ancient environments to calculate approximate temperature and humidity.

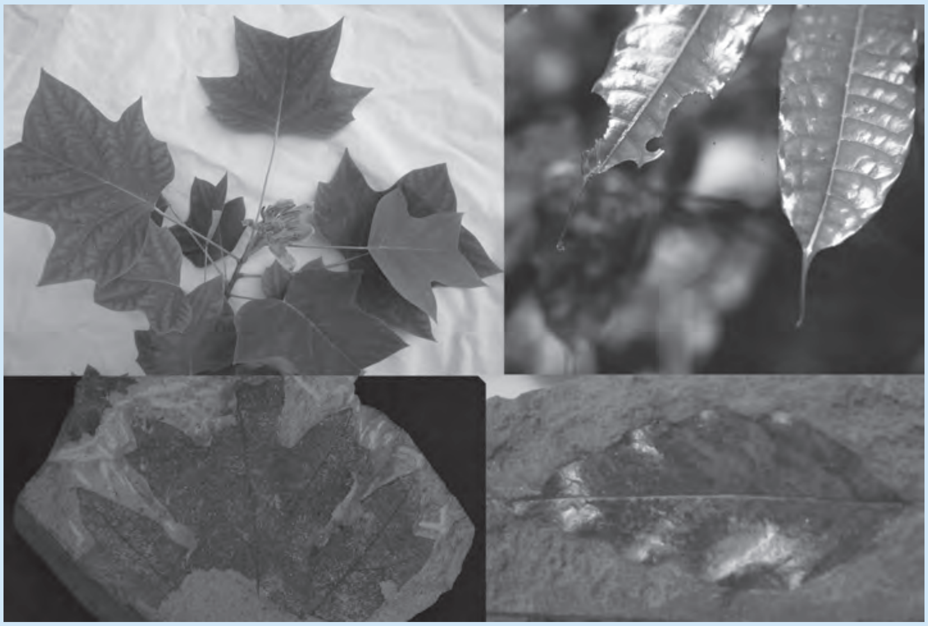


Figure B: Leaf margin shapes can be used as climate proxies. Plants with leaves with toothed or divided margins (above left) live today in cooler climates, whereas plants with leaves with smooth or entire margins (above right) live in warmer climates. This observation can be used to interpret the climates in which fossil leaves (lower left and right) grew.

Ancient plant pollen and spores (produced by plants such as ferns, lichens, and mosses) can also help us learn about ancient climates. **Palynology** (the study of pollen and spores) uses the fortunate circumstances of these objects being small, abundant, and easily preserved. Due to their tough organic coating, they are commonly preserved in the sand and sediment from places like lakes and rivers, even though trees and leaves are seldom preserved. If the pollen can be identified to a particular kind of plant, and if environmental constraints of that plant are known (by studying it or its descendants living today), the history of climate in the area can be inferred. Pollen and spores have, for example, been used to track how plant communities move north and south during fluctuations between glacial and warmer intervals.

Single-celled organisms, or **protists**, make up a large proportion of the plankton at the base of oceanic food webs. Some of these protists, especially shelled forms called foraminifera (see *Figure A* above), are particularly valuable as indicators of past climate conditions, either through analysis of the oxygen isotopes in their fossilized carbonate shells (see Box 4.4), or by comparing fossil forms to those alive today and inferring that they had similar environmental distributions.

Mesozoic • *a geologic time era that spans from 252 to 66 million years ago. This era is also called the "age of reptiles" since dinosaurs and other reptiles dominated both marine and terrestrial ecosystems. During this time, the last of the Earth's major supercontinents, PANGAEA, formed and later broke up, producing the Earth's current geography.*

leaf margin analysis • *a PROXY method for estimating past temperatures using known relationships between the shape of leaf margins (smooth or toothed) and temperature and humidity.*

palynology • *the study of modern and fossil pollen, spores, and other microscopic plant matter.*

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floodplain • the land around a river that is prone to flooding. This area can be grassy, but the sediments under the surface are usually deposits from previous floods.

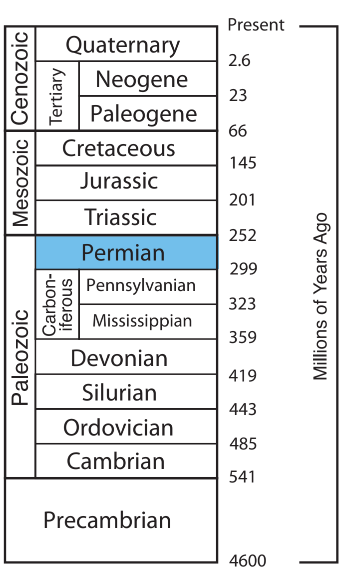
basalt • an extrusive igneous rock, and the most common rock type on the surface of the Earth. It forms the upper surface of all oceanic plates, and is the principal rock of ocean/seafloor ridges, oceanic islands, and high-volume continental eruptions. Basalt is fine-grained and mostly dark-colored, although it often weathers to reds and browns because of its high iron content.

pyroclastic flow • the rapid flow of lava, ash, and gases resulting from an explosive volcanic eruption.

Figure 4.10: Dune cross-beds in the Coconino Sandstone at Sedona, Arizona.

The continued growth of Pangaea led to a gradual shift toward a humid climate in places such as the Northwest Central U.S., where abundant, seasonal rainfall fell as intense monsoons that impacted large swaths of the continent. The climate resembled that of modern India, where monsoons soak the land in the summer and completely dry out in the winter. As the monsoon's intensity increased, the vast dune deserts of the late Permian were replaced by rivers and **floodplains**. Soils associated with these floodplains testify to the extreme seasonality of rainfall during that time.

The Permian-Triassic boundary (252 million years ago) was marked by the eruption of million km^3 of **basalt** and **pyroclastic flows**.[3] These deposits, called the Siberian Traps and found in present-day Siberia (Russia), burned through carbonate, evaporite, and organic rich sediments, which contributed to the load of greenhouse and other toxic gasses added to the atmosphere. Global shifts in carbon and oxygen isotopes preserved in the rock and fossil record indicate the widespread implications of these eruptions on the Earth's climate. Extreme warming of the ocean, possible acidification of the ocean by the dissolution of CO_2, and acid rain on the continents resulted in the largest mass extinction of the Phanerozoic, with the vast majority of marine and terrestrial faunas becoming extinct. This extinction interval is often referred to as the time when life nearly died, and the full recovery of biological diversity and the return of complex marine communities took many millions of years.



[3] The amount of rock erupted in the Siberian Traps would be enough to cover the continental US (a little over 8 million km^2) in a layer of rock half a kilometer thick!

3.4 The Mesozoic era

The Mesozoic era is known popularly as the Age of Reptiles—dominated by **dinosaurs** on land, **pterosaurs** in the air, and several groups of large marine reptiles in the ocean. Climate scientists are especially interested in the relationship between climate and mass extinctions on each end the Mesozoic, and the one at the end of the Triassic. The Cretaceous is of interest as an analog for a warm, high CO_2 world with no polar ice caps, were human-induced climate change to trigger a positive feedback loop that led, long-term, to complete melting of the Antarctic and Greenland ice sheets.

3.4.1 Triassic and Early Jurassic

By around 220 million years ago, in the mid-Triassic, what is now the U.S. moved north across the equator. Pangea began breaking up into continents that would drift toward their modern-day positions. The breakup of Pangea resulted in the development of continental **rift basins** along what is now the northeast coast of the U.S. These rift basins were filled by a string of big lakes from Virginia into Canada. One of these lakes, now called the Newark Basin, recorded in its sediments a very detailed record of climate. It shows that climate cycled annually between very wet and dry intervals, presumably connected to annual monsoons, but also over longer time periods. The record in the Newark Basin is so good and so long that we can also identify cycles occurring over tens and hundreds of thousands of years, cycles that correspond to the astronomical cycles ("**Milankovich Cycles**") that would become so influential on climate cycles of the late **Cenozoic** era.

See Chapter 3: What is Climate? for more about Milankovich Cycles.

This rifting also resulted in the eruption of extensive mantle-derived volcanic material known as the Central Atlantic Magmatic Province. These basalt deposits are preserved today in Northeastern US and Canada. This eruption, like the Siberian Traps, disrupted global climate and led to the 4th major mass extinction, of both marine and terrestrial life.

3.4.2 Jurassic

The **Jurassic** and **Cretaceous** climates remained warm, but in many areas gradually became wetter, but without the strong seasonality of the Triassic. Terrestrial environments became dominated by dinosaurs.

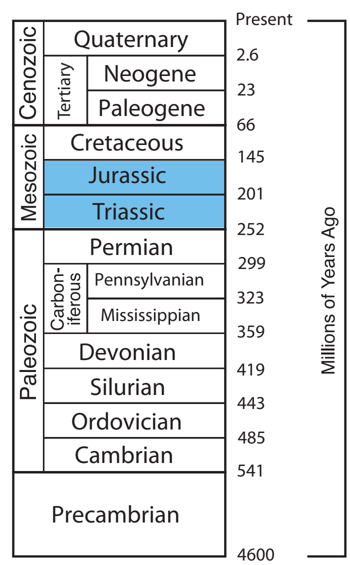
The intensity of the **monsoons** so prominent in the Triassic in the Southwestern US waned by the early Jurassic, and the rivers and floodplains of the Southwest were replaced by even larger deserts. The Southwest's Triassic-Jurassic dune deposits are some of the most extensive in the world, and the dune field that existed during the Jurassic may be the largest in Earth history. These deposits, including the Navajo Sandstone, are responsible for spectacular scenery in the national parks and recreation areas of northernmost Arizona and southern Utah (*Figure 4.11*).

Mesozoic

dinosaur • *a member of a group of terrestrial reptiles with a common ancestor and thus certain anatomical similarities, including long ankle bones and erect limbs. Most species of the large reptile groups, including the dinosaurs, disappeared at or before the MASS EXTINCTION at the end of the CRETACEOUS.*

pterosaur • *extinct flying reptiles with wingspans of up to 15 meters (49 feet). They lived during the same time as the dinosaurs.*

rift basin • *a topographic depression caused by subsidence within a rift (a break or crack in the Earth's crust); the basin, since it is at a relatively low elevation, usually contains freshwater bodies such as rivers and lakes.*



Milankovich Cycle • *cyclical changes in the amount of heat received from the Sun, associated with how the Earth's orbit, tilt, and wobble alter its position with respect to the Sun. These changes affect the global climate, most notably alterations of glacial and interglacial intervals.*

Cenozoic • *the geologic time period spanning from 66 million years ago to the present. The Cenozoic is also known as the age of mammals, since extinction of the large reptiles at the end of the MESOZOIC allowed mammals to diversify. The Cenozoic includes the Paleogene, Neogene, and Quaternary periods.*

Jurassic • *the geologic time period lasting from 201 to 145 million years ago. During the Jurassic, dinosaurs dominated the landscape and the first birds appeared. The Jurassic is the middle period of the MESOZOIC.*

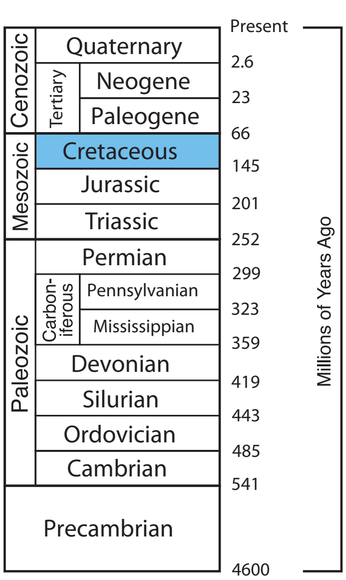


Figure 4.11: The Wave, a series of intersecting U-shaped troughs eroded into Jurassic Navajo Sandstone within the Paria Canyon-Vermilion Cliffs Wilderness, Arizona. The cycling layers in the sandstone represent changes in the direction of prevailing winds as large sand dunes migrated across the desert.

Meanwhile, the breakup of Pangea caused the Gulf of Mexico to rift open, flooding it with seawater. Because the climate was still relatively warm and dry, evaporation rates were high, and extremely thick deposits of salt accumulated there. These salt deposits have played a key role in trapping petroleum along the Gulf Coast.

Later in the Jurassic the climate of the Southwestern US became more moderate, and dune fields were replaced by rivers and floodplains populated by a rich dinosaur fauna (exemplified by the Morrison Formation). The terrestrial rocks of southeastern California contain **ginkgos** and **cycads** that indicate a warm, moderately wet climate.

3.4.3 Cretaceous

The Earth warmed near the beginning of the Cretaceous. Global temperatures were as much as 10°C (18°F) above those at present. Even though Alaska was closer to the North Pole than it is at present (*Figure 4.12*), fossil vegetation indicates that its climate was very similar to that of western Oregon today. Lush swamps and forests occupied lowland areas, and some swamps had become rich coal beds. Throughout the Cretaceous, sea level was an average of 100 meters (330 feet) higher than it is today; polar glaciers were already absent, so the increase must have been largely as a result of water displacement by rapid sea-floor spreading, such as along the mid-Atlantic Ridge as Pangea continued to split apart. Shallow seaways spread over many of the continents, and in the mid and late Cretaceous, an **inland sea**, called the **Western Interior Seaway**, divided North America in two (*Figure 4.13*). Cretaceous fossils from the Western Interior Seaway show that it supported large marine reptiles, while crocodiles and dinosaurs were abundant on land. Tropical marine fossils can be even be found as far north and inland as Minnesota. This seaway had substantial marine productivity, and its organic-rich rocks are now substantial sources of fossil fuels in the Northwest Central and Southwestern U.S.

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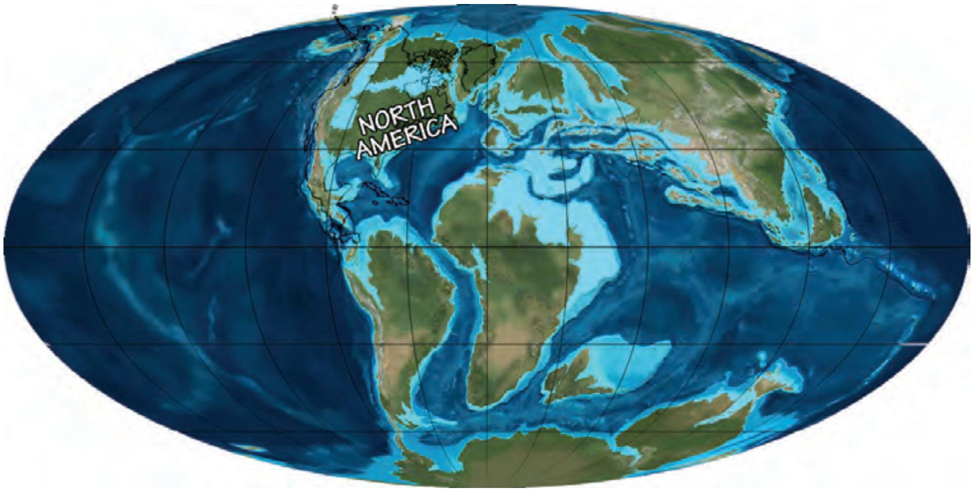


Figure 4.12: Earth during the early Cretaceous, around 105 million years ago.

Figure 4.13: The Western Interior Seaway.

3.4.4 Late Cretaceous Climate Change and End-Cretaceous Extinction

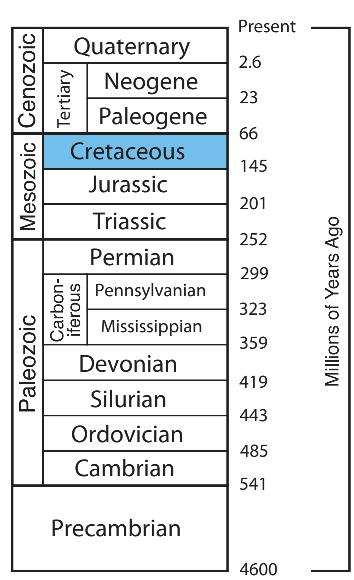
In the late Cretaceous sea level dropped. As the continents moved closer to their modern positions, global climate—though still warmer than today—cooled.

At the very end of the Cretaceous, the Gulf Coast experienced an enormous disruption when a large asteroid or **bolide** collided with Earth in what is now the northern Yucatán Peninsula in Mexico. The impact vaporized both water and rock, blocking out sunlight for weeks to years, which led to a collapse

Mesozoic

monsoon • *a seasonal wind pattern in the Indian Ocean and South Asia which reverses direction between southwesterly and northeasterly, creating a wet season in summer and a dry season in winter.*

ginkgo • *a terrestrial tree belonging to the plant division Ginkgophyta, and characterized by broad fan-shaped leaves, large seeds without protective coatings, and no flowers. Ginkgos were very common and diverse in the Mesozoic, but today only one species exists, Ginkgo biloba.*



of photosynthesis and food webs on land and in the oceans. These factors resulted in the 5th and most mass extinction of the Phanerozoic (other than the one we may be in the middle of today.[4] This event famously led to the demise of non-avian dinosaurs,[5] marine reptiles, and many invertebrates such **ammonoids**. After this event, the climate may have cooled briefly, but it soon rebounded to a warmer state.

3.5 Cenozoic

The Cenozoic era started out warming, but ultimately was overall time of cooling, starting with developing of ice sheets in Antarctica and leading to Quaternary **glacial-interglacial cycles**. The interval is also the time over which modern ecosystems developed, dominated on land by mammals, birds, and flowering plants. The Cenozoic contains a diversity of climate analogs that climate scientists find useful because the position of continents and nature of the climate system is relatively similar to today.

3.5.1 Paleocene and Eocene

Climate warmed during the **Paleocene**, culminating at the boundary between the Paleocene and **Eocene** epochs (around 56 million years ago) with temperature spiking suddenly upward. Geologists call this the Paleocene-Eocene Thermal Maximum (PETM). During the warming event the atmosphere and ocean warmed by as much as 8°C (14°F) in as little as 4000 years, and deep oceans became acidic, resulting in the dissolution and extinction of shelly marine animals. The causes of this event remain unclear, but may have involved the sudden release of methane from sediments on the seafloor. The resulting greenhouse effect persisted for 100,000 years. The PETM is of great interest to climate scientists because it is in some respects the most similar analog to rapid increases in greenhouse gases that we are currently experiencing.

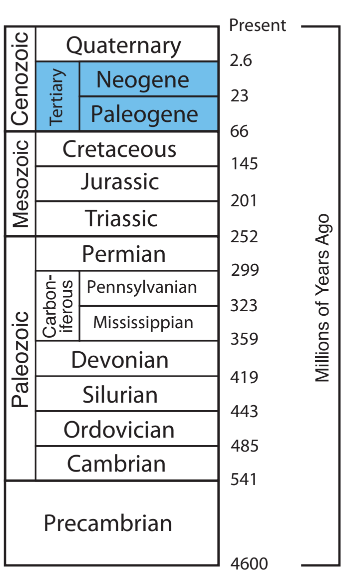
During the Eocene the climate remained relatively warm, with palm trees growing in southern Alaska. Records of plants and animals found in Oregon and Washington indicate that the northwestern US was home to a subtropical rainforest with (depending on the site) banana and citrus trees, palm trees, ferns, and dawn redwoods. The Southwest's climate was warm and wet, with strong volcanic activity. Large lakes covered parts of northern Utah, Colorado, and Wyoming (the Greater Green River Basin) (*Figure 4.14*). Climates were warm enough for crocodiles to live as far north as 50°N in the interior of North America and on Ellsemere Island of Northern Canada around 78°N. Warm climate are also reflected in the land plants and diversity of marine life, for example, in the rich fossil record of clams, snails, and echinoderms found in the Gulf and Southeast Coastal Plain.

During the Eocene, India began to collide with Asia to form the Himalayas. The formation of the Himalayas over the span of tens of millions of years had a

cycad • a palm-like, terrestrial seed plant (tree) belonging to the Class Cycadopsida, and characterized by a woody trunk, a crown of stiff evergreen leaves, seeds without protective coatings, and no flowers. Cycads were very common in the MESOZOIC, but are much reduced in diversity today, restricted to the tropical and subtropical regions of the planet.

inland sea • a shallow sea covering the central area of a continent during periods of high sea level. An inland sea is located on continental crust, while other seas are located on oceanic crust.

Western Interior Seaway • an INLAND SEA which divided North America in two along a North-South axis during the mid and late CRETACEOUS.



[4] There is an extensive literature estimating modern rates of extinction, comparing them to mass extinctions of the geologic past, and estimating the role climate change may have in future extinctions. Elizabeth Kolbert's book The Sixth Extinction: An Unnatural History (Henry Holt and Co.: NY, 336 pp.) is a very readable introduction to the topic.

[5] Because birds evolved from dinosaurs, they are technically considered to be dinosaurs. Thus, for clarity, paleontologists use the term "non-avian" dinosaurs to refer to all dinosaurs that are not birds.

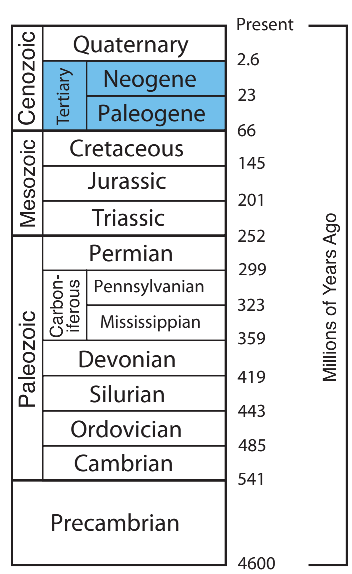
Figure 4.14: Well-preserved fossils from the Green River Formation, southwestern Wyoming. A) Palm frond, Sabalites powelli, *about 1.2 meters (4 feet) long, with fossil fish* Knightia. *B) An undetermined bird species with preserved feathers, about 25 centimeters (10 inches) long. C)* Heliobatis radians, *a stingray, about 40 centimeters (16 inches) long, with fossil fish. D)* Borealosuchus wilsoni, *a crocodilian, reached lengths of 4.5 meters (15 feet).*

significant impact on global climate, with the chemical weathering of the newly exposed rock serving as a sink for atmospheric CO_2. With the reduction of this greenhouse gas, global temperatures began cooling, the start of a long downward trend through much of the remainder of the Cenozoic.

bolide • *an extraterrestrial object of any composition that forms a large crater upon impact with the Earth. In astronomy, bolides are bright meteors (also known as fireballs) that explode as they pass through the Earth's atmosphere.*

ammonoid • *a group of extinct cephalopods belonging to the Phylum Mollusca, and possessing a spiraling, tightly-coiled shell characterized by ridges, or septa.*

glacial-interglacial cycle • *an alternation between times in Earth's history when continental ICE SHEETS grow and advance toward lower latitudes (GLACIALS), and times when the climate is warmer and ice sheets melt back (INTERGLACIALS).*



3.5.2 Oligocene, Miocene, and Pliocene

Global temperatures fell sharply at the boundary between the Eocene and **Oligocene** epochs (around 35 million years ago), due in part to the separation of South America's southern tip from Antarctica. This allowed for the formation of the Antarctic Circumpolar Current, which insulated Antarctica from warm ocean water coming from lower latitudes. Antarctica moved south, and by 30 million years ago, temperatures were low enough that glaciers began to grow on its mountains. An "ice age" can be described as the presence of long-term high-latitude glaciers, and by this broad definition, the current (today's) ice age began over 30 million years ago with the appearance of ice sheets on Antarctica.

Between 35 and 20 million years ago the climate in the Western U.S. became cooler and drier, and prairies and deciduous trees such as oak, maple, and alder flourished. On the Great Plains, grasses, and mammals specialized to feed on them, increased in prominence as the **Miocene** became drier. This coincided with the initial uplift of the Cascade Range (37–7 million years ago), which began to create a rain shadow to the east. The final uplift of the Cascades and Sierra Nevada created the intense rain shadow that is responsible for the aridity of eastern Washington, eastern Oregon, and Nevada today.

Global temperatures fell further in the Miocene as the Himalayas continued to grow and weather, serving as a sink removing CO_2 from the atmosphere. With the reduction of this greenhouse gas, temperatures cooled worldwide, and this cooling continued more-or-less into the **Pleistocene**. By about 15 million years ago ice covered much of Antarctica and had begun to form on Greenland. As high latitude glaciers grew, sea levels dropped.

In the mid-Miocene, especially around 17 to 14 million years ago, eruptions in eastern Oregon produced enormous amounts of basalt that flowed north and west, filling the Columbia River basin. These are some of the largest such eruptions in the history of the Earth, and they continued over a span of about 11 million year, finally ceasing about 6 million years ago. While evidence that these eruptions influenced global climate is ambiguous, climatic changes are recorded in soils that formed atop some of the lava flows. These soils indicate a decrease in temperature after a period known as the Middle Miocene Climatic Optimum, a brief warming episode that occurred around 16 million years ago. Miocene warming is reflected in the diverse marine and terrestrial fossils of the Atlantic Coastal Plain, which extends from Maryland to Florida.

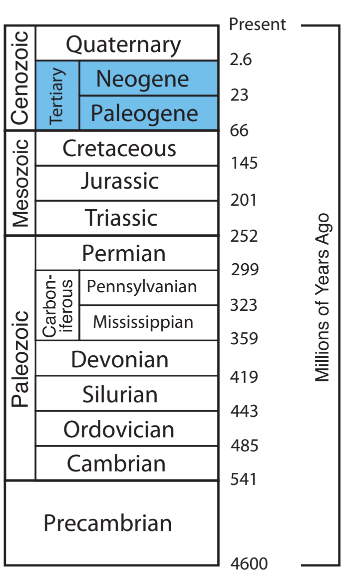
Around 3.5 million years ago, glacial ice began to form over the Arctic Ocean and on the northern parts of North America and Eurasia (*Figure 4.15*). Surprisingly, a major contributing factor to this event was a geological change that occurred half a world away. The Central American Isthmus, which today makes up most of Panama and Costa Rica, rose out of the ocean at around this time, formed by undersea volcanoes. The new dry-land isthmus blocked the warm ocean currents that had been flowing east-to-west from the Atlantic to the Pacific for more than 100 million years, diverting them into the Gulf of Mexico and ultimately into the western Atlantic Gulf Stream. The strengthened Gulf Stream carried more warm, moist air with it into the northern Atlantic, which caused increased snowfall in high latitudes, leading to enhanced glacier development and accelerating cooling. Such changes contributed to the high

Paleocene • first geologic time interval of the Cenozoic era, spanning from about 66 to 56 million years ago.

Eocene • a geologic time interval extending from 56 to 33 million years ago. The Eocene is the second epoch of the Cenozoic era.

Oligocene • third geologic time epoch of the Cenozoic era, spanning from about 34 to 23 million years ago.

Miocene • fourth geologic epoch of the Cenozoic era, extending from 23 to 5 million years ago.



Pleistocene - an epoch of the QUATERNARY period, lasting from 2.6 million to about 11,700 years ago. During the Pleistocene, continental ice sheets advanced south and retreated north several dozen times.

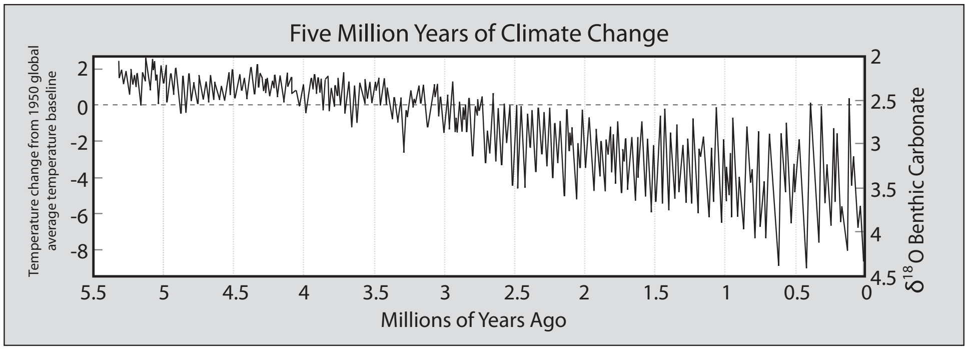


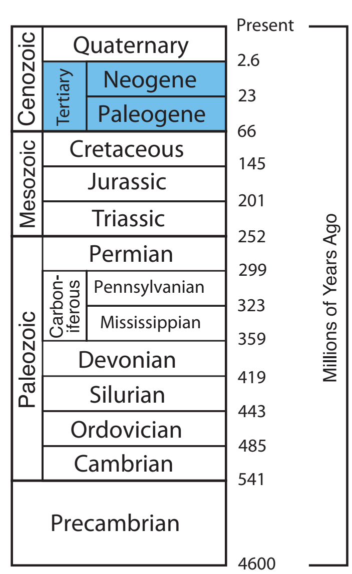
Figure 4.15: Five million years of climate change. In the graph, oxygen isotopes from fossil shells of deep-sea marine organisms (foraminifera) have been used to show relative global temperatures for the last 5.5 million years. The curve is influenced both by the amount of water stored in ice sheets and by temperature at the bottom of the ocean; both cooler temperatures and larger ice sheets cause higher $\delta^{18}O$ values. On the left vertical axis are proxy temperature data from an ice core. See Box 4.4 *for more about using oxygen isotopes.*

Box 4.4: Using oxygen isotopes to determine past climates

The different chemical elements, like oxygen, carbon, and hydrogen that we encounter in the Periodic Table in chemistry class are distinguished by their differing numbers of subatomic particles: each element has a distinct number of protons, and an equal number of electrons. Isotopes are variants of elements that have the same numbers of protons and electrons, but differ in the number of neutrons. This means that different isotopes of an element have a slightly different mass.

The most common isotope of oxygen has 16 neutrons and is therefore called oxygen 16, abbreviated ^{16}O. A small proportion of the oxygen in the universe has 18 neutrons; oxygen 18 (^{18}O). Because ^{16}O has fewer neutrons than ^{18}O, it behaves differently. For example, it is more easily integrated into water vapor, and so clouds and their associated precipitation contains relatively more ^{16}O than the lake or ocean from which the water evaporated. When this precipitation is stored for a long time in the form of compacted snow in glaciers, as a result of colder climate, the oceans of the world have relatively less ^{16}O in their water than they do in warmer times. We call oceans that are enriched in ^{18}O, during these glacial intervals, isotopically “heavy” because they contain more of the neutron rich oxygen 18.

^{16}O is also more easily incorporated into chemical compounds. Many marine organisms make their shells out of calcium carbonate ($CaCO_3$), and need to take dissolved carbonate ions out of the seawater to do this. Therefore, when they build their shells, marine organisms record the proportion of ^{16}O that exists in seawater at the time. Because of the different behavior of the two isotopes of oxygen, shells have a higher proportion of ^{16}O in a warmer climate when the lighter isotope of oxygen is more prevalent in ocean water and not stored in glaciers. When the shells are preserved as fossils on the sea floor, and then extracted in a sediment core, they can be analyzed for their amount of ^{16}O relative to their amount of ^{18}O to estimate ancient temperatures. Scientists commonly use the quantity $\delta^{18}O$ (pronounced "delta-18-oh"), which reflects the ratio of ^{18}O to ^{16}O compared to a standard; smaller values of $\delta^{18}O$ indicate higher temperatures (e.g., in *Figures 4.6* and *4.15*).



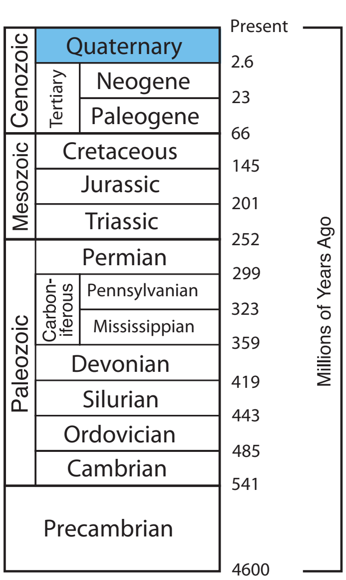
Cenozoic

Pliocene • fifth geologic epoch of the Cenozoic era, extending from roughly 5 to 2.5 million years ago.

Quaternary • a geologic time period that extends from 2.6 million years ago to the present. This period is largely defined by the periodic advance and retreat of continental glaciers. The Quaternary is part of the CENOZOIC.

glacial • a time in Earth's history when a cold climate leads to the advance of GLACIERS and ICE SHEETS. See also INTERGLACIAL.

interglacial • a time in Earth's history between GLACIAL advances; there have been about 50 glacial advances and interglacials in the past 2.5 million years.



amplitude glacial-interglacial cycles of the Pleistocene. These changes in ocean circulation throughout the Caribbean and Gulf of Mexico also affected nutrient supplies in the coastal ocean, which may have contributed to an increase in the extinction of marine animals (including everything from mollusks and corals to whales and dugongs) during the late **Pliocene**.

3.5.3 Pleistocene

The start of the **Quaternary** Period, and the Pleistocene Epoch, are defined by a global drop in Earth's temperatures as recorded by ice and ocean sediment records (see *Box 4.7*). A sheet of sea-ice formed over the Arctic, and ice sheets spread over northern Asia, Europe, and North America, as the most recent "Ice Age" took hold. Ice sheets have advanced and retreated dozens of times over the past 2.6 million years (see *Box 4.5*), controlled by variations in the Earth's orbit, rotational tilt, and relative amount of wobble around its rotational axis (See *Box 4.6* on Milankovitch cycles). Since each glacial advance scrapes away rock and reworks the geologic evidence of previous glacial events, it can be difficult to reconstruct the precise course of events. Therefore, to investigate the details of any associated climate change we must seek environments that record climate change and are preserved in the geologic record. Since the 1970s, the international Deep Sea Drilling Project has provided a treasure trove of data on coincident changes in the ocean, preserved in sediments at the ocean bottom (*Figure 4.16*). In the 1980s, coring of ice sheets in Greenland and Antarctica provided similar high resolution data on atmospheric composition and temperature back nearly one million years (*Figure 4.17*). The data from these programs have revealed that the Earth experienced dozens of warming and cooling cycles over the course of the Quaternary period (the past 2.6 million years). Traces of the earlier and less extensive Pleistocene glacial advances that must have occurred have been completely erased on land, so these advances were unknown before records from deep-sea cores and ice cores revealed them.

Chemical, sedimentological, and marine organism data has enabled researchers to compile an extensive and precise record of changes in global ice volume and thus glacial advances, and to make sense of the glacial cycles in terms of orbital variations of the Earth around the sun. These orbital variations, called Milankovitch cycles (see *Box 4.6*), result in changes in incoming solar radiation (insolation). While they occurred throughout Earth's history, no matter the average global temperature, they have an especially large impact during cooler intervals of Earth's history where ice forms at the poles. When Earth was relatively warm, these orbital variations most notably caused changes in precipitation. When the Earth became cooler, however, as happened approximately 2.6 million years ago, the orbital variations resulted in changes in global temperature. Thus, roughly every 40,000 years from 2.6 million to 0.7 million years ago, and every 100,000 years since 0.7 million years ago, ice sheets have expanded into lower latitudes, at their greatest extent reaching the northern parts of what is now the United States. Scientists call these expansions **glacials**, the last one peaking approximately 20,000 years ago in what is called the Last Glacial Maximum. The warmer intervals between glacials, when the ice retreated northward, are called **interglacials**. Earth is currently in an interglacial interval. Prior to the present, the most recent interglacial period occurred approximately 125,000 years ago when temperatures at the poles were 3-5 degrees C warmer than at present, and global sea level was 4-6 meters (13-20 feet) higher than it is today.

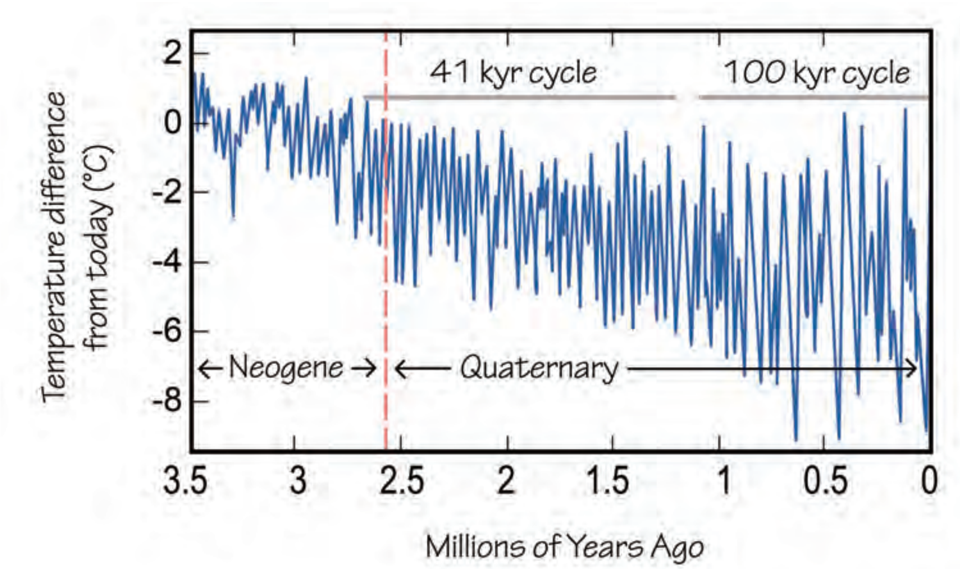


Figure 4.16: Ocean bottom temperatures from 3.6 million years ago to present, based on chemical analyses of foraminifera shells. Notice how the amplitude of glacial-interglacial variations increases through time, and how the length of cycles changes.

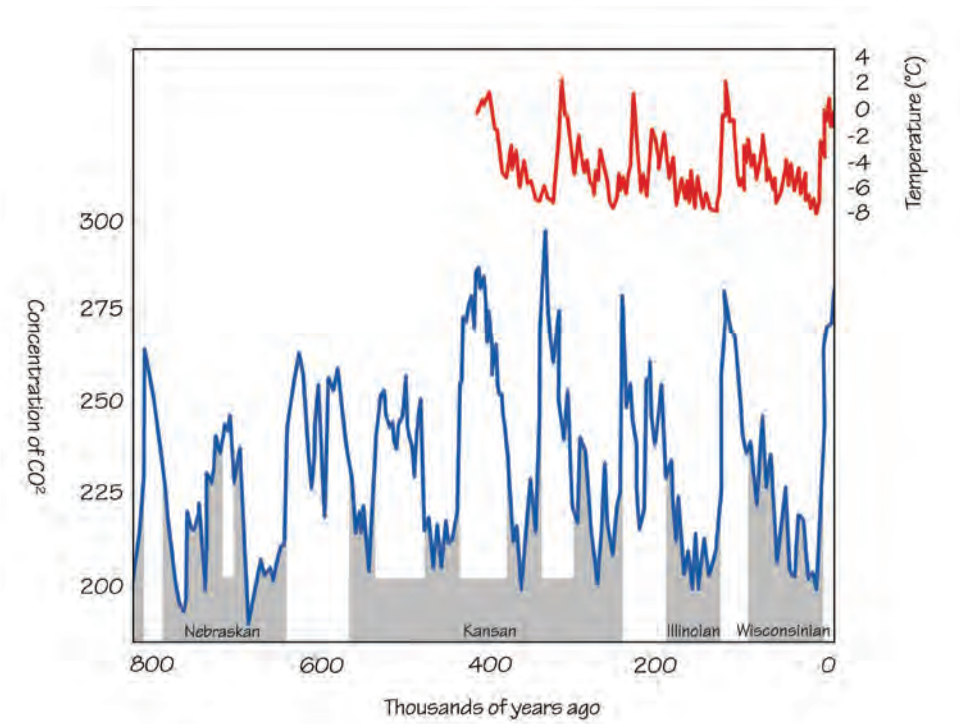
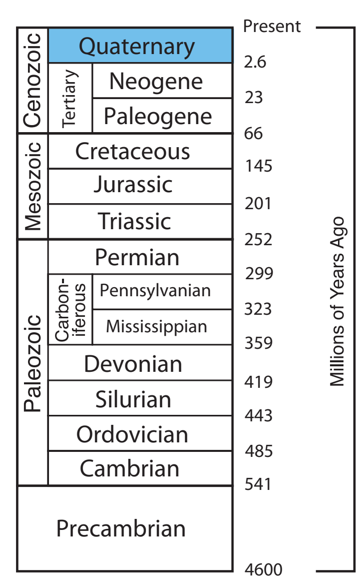


Figure 4.17: Ice core atmospheric temperature and carbon dioxide concentrations from an ice core taken in Vostok in Antarctica along with CO_2 data from several cores in Greenland give a record of glacial advances over the past 800,000 years. Note that Kansan and Nebraskan deposits represent more than one glacial advance.



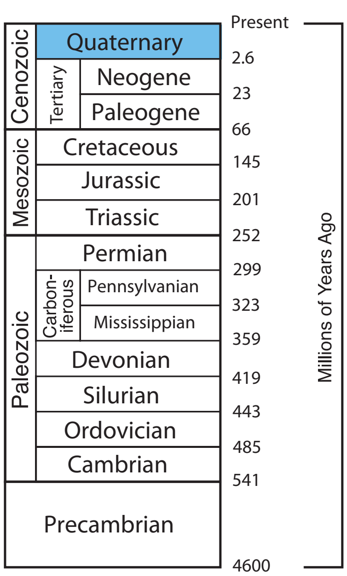
Box 4.5: Age of the Quaternary

In 2009, scientists at the International Commission on Stratigraphy voted to move the beginning of the Quaternary period to 2.6 million years ago, shifting it 0.8 million years earlier than the previous date of 1.8 million years ago—a date set in 1985. They argued that the previous start date was based on data that reflected climatic cooling that was only local to the region in Italy where it was first observed. In contrast, the 2.6-million-year mark shows a global drop in temperature, and it includes the entirety of North American and Eurasian glaciation, rather than having it divided between the Quaternary and the earlier Neogene period.

Box 4.6: Astronomic cycles and ice sheets

The cyclical movements of ice sheets seem primarily to be caused by specific astronomic cycles called Milankovitch cycles, which change the amount of light the Earth receives, particularly when comparing the summer to the winter. The cycles, predicted through principles of physics a century ago, are related to the Earth's eccentricity, or the shape of Earth's orbit around the sun which varies on 100,000 year time scales, the degree of tilt of the Earth, which varies on 41,000 year cycles, and the precession, or wobble of Earth as it rotates over periods of 23,000 years. When the cycles interact such that there is milder seasonality (cooler summers and warmer winters) at high latitudes in the Northern Hemisphere, less snow melts in summer, which allows glaciers to grow. The cyclicity of glacial-interglacial advances was about 40,000 years from before the start of the Quaternary until about a million years ago, likely controlled by Earth's rotational angle. For reasons that aren't clear, however, the cycles changed to about 100,000 years, controlled more by the eccentricity of Earth's orbit. If not for human-induced climate change, we might expect glaciers to approach Kansas and Missouri again in about 80,000 years.

For graphics and more detailed information about the mechanisms of Milankovich Cycles, see Chapter 3: What is Climate?



The continental glaciers that repeatedly covered parts of North America during the Quaternary had their origin in northern Canada. As the climate cooled, more snow fell in the winter than melted in the summer, causing the snow to pack into dense glacial ice. As more snow and ice accumulated on the glacier (and less melted), the ice began to move under its own weight and pressure. The older ice on the bottom was pushed out horizontally by the weight of the overlying younger ice and snow. Glacial ice then radiated out from a central point, flowing laterally in every direction away from the origin (*Figure 4.18*). And thus, a continental glacier originating in far northern Canada began to move south towards the Northeastern U.S. (*Figure 4.19*). The ice sheet crept slowly forward, scraping off the loose rock material and gouging the bedrock beneath as it advanced. Glaciers stop growing when the rate of flow of the glacier from the north is offset by the rate of melting along the southern edge of the glacier.

Using bubbles and water trapped in ice cores (see *Box 4.7*), scientists can measure the past atmospheric concentration of CO_2. Atmospheric CO_2 co-

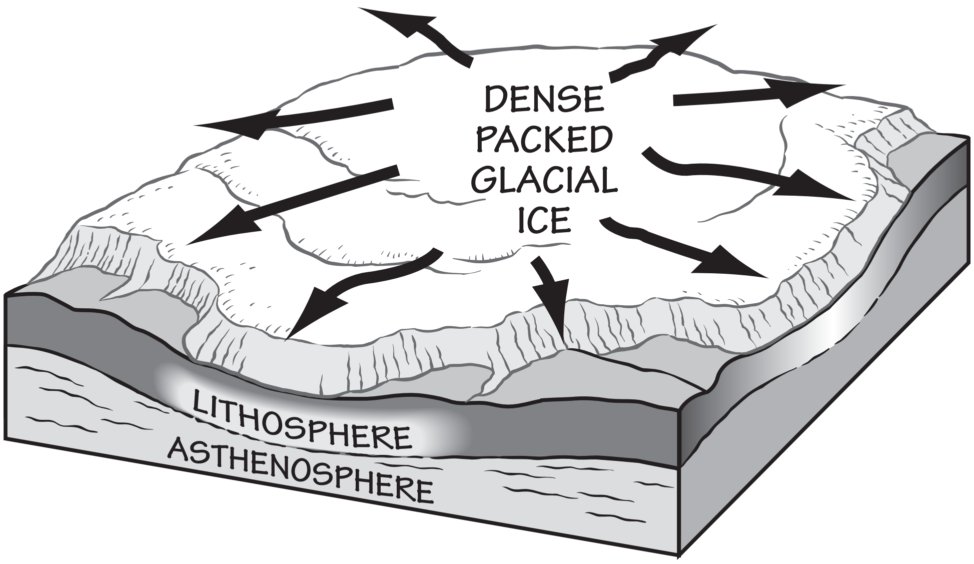


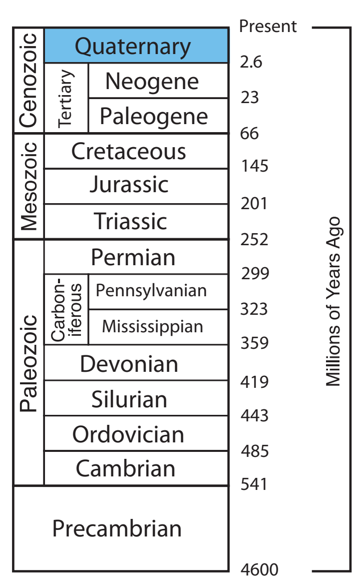
Figure 4.18: As ice piles up, a glacier forms, then begins to flow outward under its own weight and pressure.

Figure 4.19: A simplified version of the advancement of the most recent ice sheet over North America.

varies with temperature in a way that reflects a positive feedback (*Figure 4.20*). For example, transitions from a cooling (glaciation) interval to a warming (deglaciation) interval occur due to changes in solar insolation associated with **Milankovitch cycles**; the warming in turn leads to release of CO_2 from warming ocean water and uncovered soils, increasing atmospheric CO_2 concentrations and further contributing to warming. The reverse occurs in transitions from warm intervals to cool intervals.

The most extensive glacial advances were recognized by their moraines long before we could date them precisely or knew the total number of glacial advances from isotopes in cores. In North America, these glacial-interglacial cycles are known as the pre-Illinoian (1.8 million to 302,000 years ago), Illinoian (191,000–131,000 years ago), Sangamonian (131,000–85,000 years ago), and

Milankovitch cycles - cyclical changes in the amount of heat received from the Sun, associated with how the Earth's orbit, tilt, and wobble alter its position with respect to the Sun. These changes affect the global climate, most notably alterations of glacial and interglacial intervals.



Holocene – the most recent portion of the QUATERNARY, beginning about 11,700 years ago and continuing to the present. It is the most recent (and current) interglacial, an interval of glacial retreat. The Holocene also encompasses the global growth and impact of the human species.

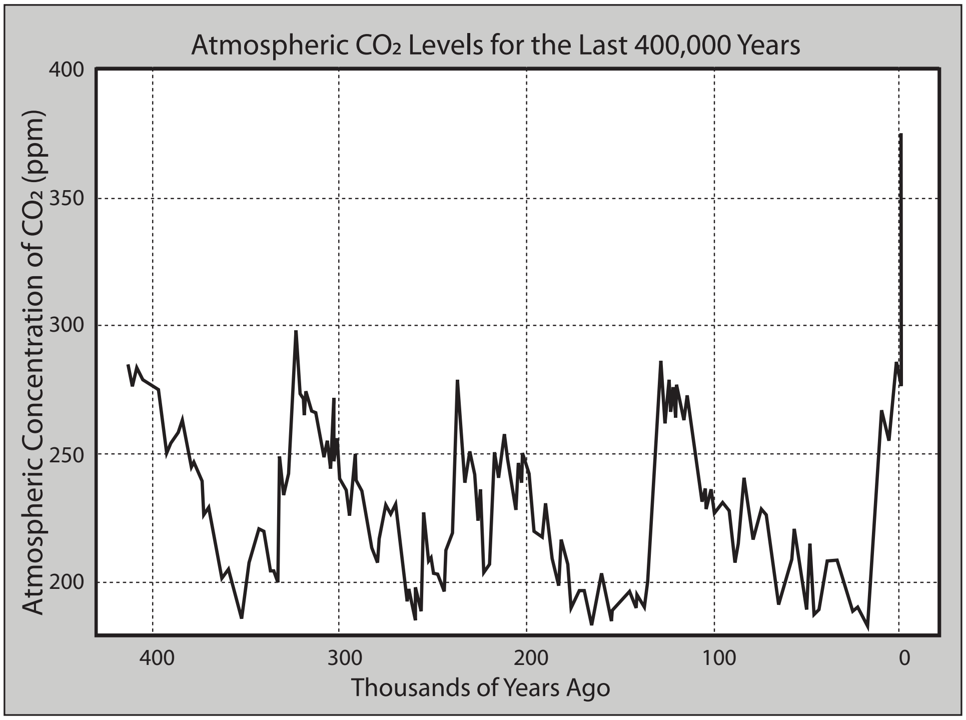


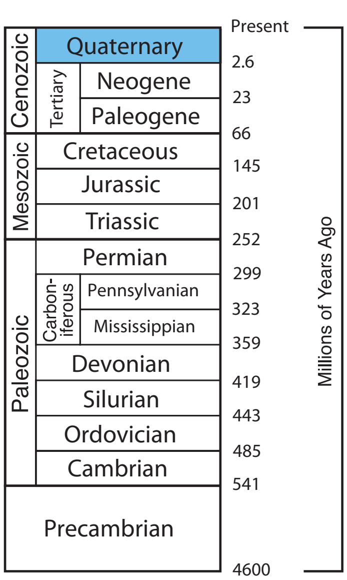
Figure 4.20: Atmospheric CO_2 during the last 400,000 years. Atmospheric CO_2 concentration has been recorded for almost 800,000 years from ice-core air-bubble samples. During that interval, including the last 450,000 years depicted on this graph, the atmospheric concentration of CO_2 has never gone above 350 parts per million until the 1980s, about 100 years after the start of the Industrial Revolution.

Wisconsinan (85,000–11,000 years ago). The Illinoian and Wisconsinian were cooler periods that saw glaciers advance, while the Sangamonian was a warm interglacial period.

The pre-Illinoian glaciation included many glacial and interglacial periods that were once subdivided into the Nebraskan, Aftonian, Kansan, and Yarmouthian ages. New data and numerical age dates suggest that the deposits are considerably more complicated; they are now lumped together into a single interval. The problem is that it can be very difficult to date glacial deposits precisely, and glacial advances wipe out much of the evidence of the previous advance. Today study of glacial cycles focuses on deep sea and ice cores, where oxygen isotopes can be measured and correlated with numbered isotope stages.

Ice sheets have come as far south as northern Missouri and northeastern Kansas, but the ice sheets did not extend into southern U.S. states, even at their largest. However, there were influences even on areas without ice sheets. For example, mountain glaciers were at high elevations in the Southwestern US, and large lakes formed in low areas. Much of the Mississippi River's great delta and alluvial fan was deposited when the glacial ice melted, creating rivers that eroded older rocks as well as carrying sediments previously scoured by the glaciers.

There are countless examples of distributions of organisms that show that temperatures in southern parts of the U.S. were cool by comparison with **Holocene** temperatures. For example, in southern New Mexico, Pleistocene fossil mammals are found that now live at higher elevations in the mountains



Box 4.7: Proxies from ice cores

In a few cases, scientists can sample ancient atmospheres directly. Ice sheets and glaciers, which can be hundreds or thousands of feet thick, are formed from snow that has collected each year on the surface, has been compressed by overlying snow and ice for many years, and ultimately recrystallizes into thick glacial ice. Bubbles in this ice can contain air that was trapped when the ice formed. The chemical composition of the air in these bubbles, as well as the frozen water surrounding them, can reveal, for example, the amount of CO_2 in the ancient atmosphere.

An **ice core** (*Figure*) is a large cylinder of ice extracted from an ice sheet or glacier, such as is found in Antarctica, Greenland, or on very high mountains worldwide. To collect ice cores, scientists use a hollow drill a few inches wide that cuts around a central cylinder of ice. Drillers carry out many cycles of lowering the drill, cutting a limited section (usually 4–6 m long), then raising all the equipment to the surface, removing the core section and beginning the process again. Much care is taken to ensure that the core is uncontaminated by modern air and water. The core is stored in an airtight plastic bag as soon as it reaches the surface, and analyzed only in a "clean room" designed to prevent contamination. To keep the ice core from degrading, it is kept well below freezing, usually below -15°C (5°F).

Ice cores record history. They can tell scientists about temperature, ocean volume, rainfall amount, levels of CO_2 and other gases in the atmosphere, solar variability, and sea-surface productivity at the time that the ice formed. Scientists can conduct chemical and isotopic studies on the ice itself, but they can also physically look at inclusions in the ice, like wind-blown dust, ash, or radioactive substances that can tell us about the extent of deserts, volcanic eruptions, forest fires, and even meteor impacts. The length of the record is extremely variable. Some cores only record the last few hundred years, whereas the longest core ever taken (from Vostok Research Station, Antarctica) allows study of climate change for over 400,000 years. Compiling information from multiple cores, scientists have now assembled a climate history of almost 800,000 years.

An ice core from Vostok Research Station, Antarctica, at the National Ice Core Laboratory in Denver, Colorado.

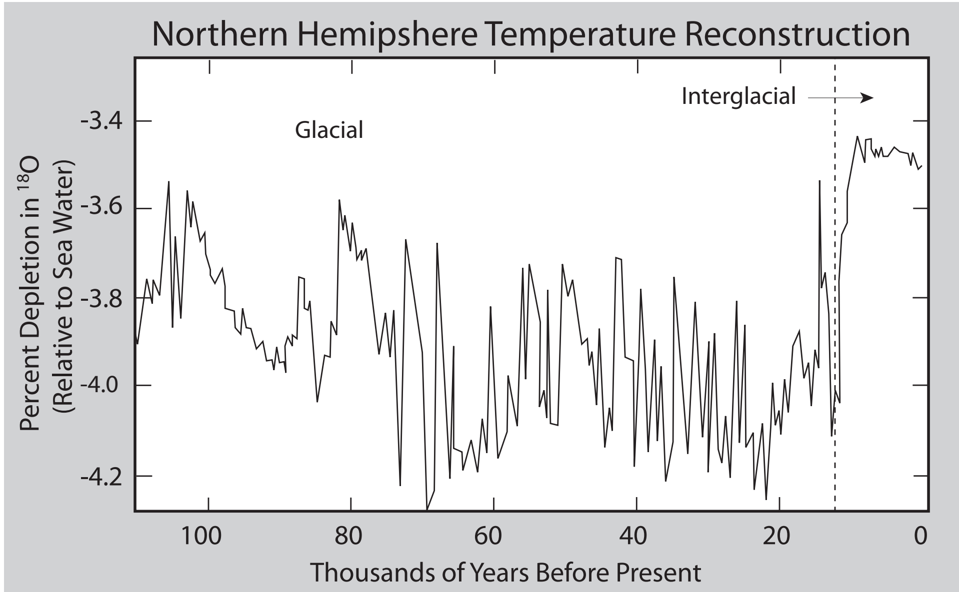


Figure 4.21: Greenland ice-core proxy temperature. Temperature records from the Green¬land Ice Core based on $\delta^{18}O$ relative to $\delta^{16}O$ for the last 100,000 years.

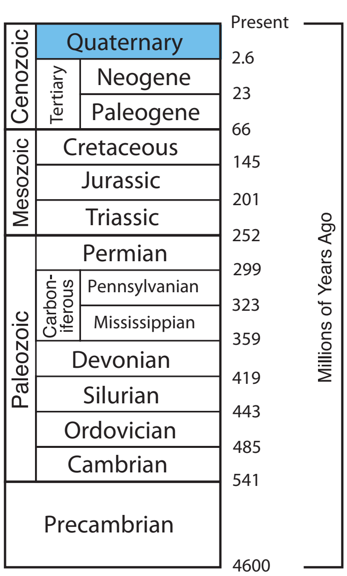
of northern New Mexico, indicating cooler temperatures and more available moisture in the area during the late Pleistocene. California's coastal climate was cooler: for example, microfossil evidence from the Rancho La Brea Tar Pits in Los Angeles tells us that southern California's climate around 40,000 years ago was similar to San Francisco's today.

3.5.4 Most Recent Glacial-Interglacial Cycle (110,000–10,000 Years Ago)

Using oxygen isotopes from ice cores (see *Box 4.7*), scientists can reconstruct Earth's global temperature during the last 100,000-year cycle (*Figure 4.21*). Tracing the midpoint of all of the high and low values in the line shows that temperature slowly dropped over this time. Geologically, scientists equate this cooling with the most recent glacial interval, when mastodons and mammoths roamed North America and a thick sheet of ice covered places like New York, Michigan, and northern Europe. The most significant recent advance of glacial ice over North America—the Laurentide Ice Sheet—peaked about 20,000 years ago, known as the "Last Glacial Maximum."

The ice sheet's maximum extent reached into Washington, Montana, the Dakotas, and Nebraska (*Figure 4.22*); it extended to where Chicago is now located, covering the northern half of the Midwestern US; and to the east as far as northern Pennsylvania and across to Long Island. The temperatures in areas not then covered in ice were moderated by its presence: the portions of the Northern US that were not covered by ice experienced a variety of cold climates and abundant lakes. The ice sheet did not, however, extend northwest into central or northern Alaska since the local climate was very dry, though an ice cap covered the Brooks Range.

After the Last Glacial Maximum ice then began melting back as global temperatures relatively abruptly increased. The Laurentide Ice Sheet and alpine ice caps throughout the Rockies melted back, leaving behind rugged



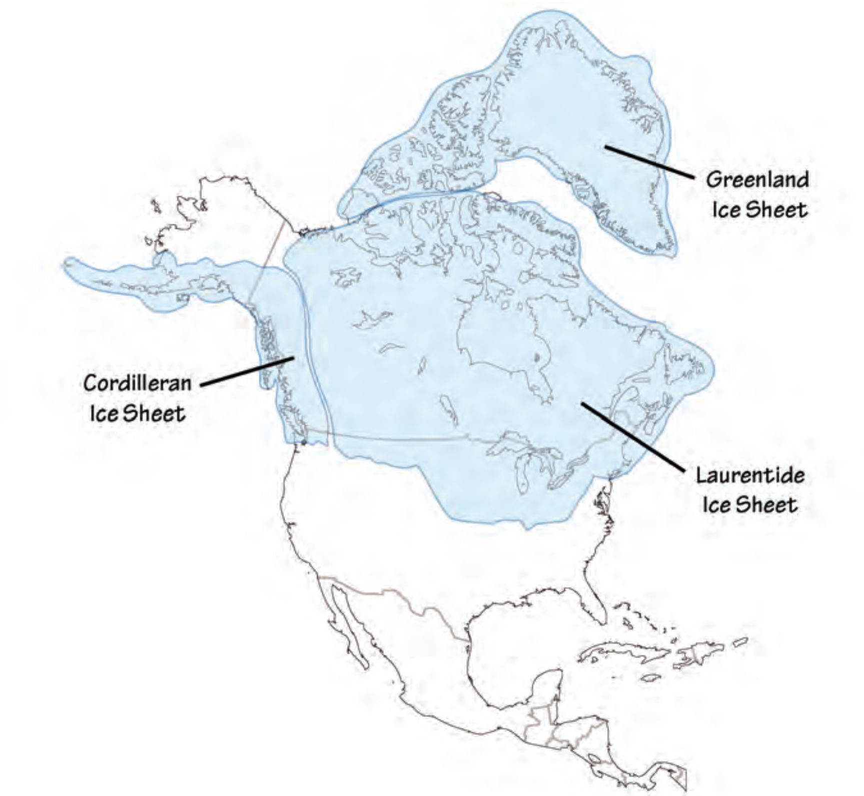


Figure 4.22: Extent of glaciation over North America during the last glacial maximum.

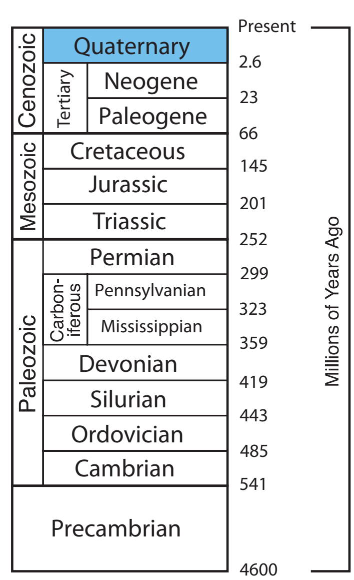
esker • a sinuous, elongated ridge of sand and gravel. Most eskers formed within ice-walled tunnels carved by streams flowing beneath a glacier. After the ice melted away, the stream deposits remained as long winding ridges. Eskers are sometimes mined for their well-sorted sand and gravel.

kettle • a depression formed where a large, isolated block of ice became separated from the retreating ICE SHEET. The ice becomes buried by sediment; when it melts, it leaves a shallow depression in the landscape that often persists as a small lake.

mountain ranges, deep glacial valleys, and plains covered with thick deposits of glacial sediment. Glaciers left behind many geologic features that define the landscape of northern parts of the U.S., including **eskers**, **kettles**, and thick deposits of sand and gravel. They scraped the surface and left behind sediment that filled uneven surfaces, having the overall effect of smoothing the landscape. But they also left behind basins carved by glaciers such as the Great Lakes, and the Finger Lakes in New York State. The northern U.S. owes a large share of its present topography and drainage patterns to the last glacial advance and meltback.

Post-glacial lakes in the Northwest Central U.S. were a part of two famously large flooding events, among the largest floods on Earth. The first was the Bonneville megaflood: melting glaciers fed the waters of ancient Lake Bonneville (the remains of which is today the Great Salt Lake), which broke through a dam of loose sediment and rapidly drained northward through southern Idaho, along what is now the Snake River, all the way to northern Idaho.

The second was a series of floods that occurred when the ice sheet alternately blocked and retreated from what is now the Clark Fork River in northwestern Montana and northern Idaho. When the river was blocked, an enormous lake built behind the ice dam, and when the ice dam failed, the water was released catastrophically. Although the floods mostly affected central Washington, large ripples from the intense flow are preserved both near Missoula, Montana, and just downstream from where the ice dammed the river in northern Idaho. These floods cut through the dust deposits and basalt that covered much of the region, leaving islands, escarpments, and channels so large that ground-based geologists did not at first recognize their origins.



Medieval Warm Period – a period of warm climate in the North Atlantic region during approximately the years 950 to 1100.

With glacial retreats, vegetation gradually changed from cold-adapted taxa along the margins of the ice sheet to the flora we see today. For example, around 12,000 years ago, New England's climate was much like that of modern-day northern Canada. Temperatures increased with time (though interrupted by a cooling period from 10,800 to 10,000 years ago called the Younger Dryas). Pollen records from New York, Connecticut, and Maine indicate a landscape dominated by boreal forests during this time, with trees such as spruce, fir, and birch. Between 13,000 and 8500 years ago, fossil evidence shows that spruce and aspen forests grew in areas of North Dakota that are now warmer, drier, and covered with prairie. Idaho became more humid and warmer than it was during the last glacial maximum.

If the past is predictor of the future, then without human intervention, global average temperature and atmospheric levels of CO_2 should slowly decrease for the next 60,000 to 80,000 years.

3.5.5 The Holocene: The Last 10,000 Years

The last 10,000 years or so look dramatically different from the variation in temperature of the previous glacial interval; there has been much less variation. Presently, the continental ice sheets and ice caps of the Pleistocene are gone, but some 150,000 alpine glaciers remain worldwide, and this time from the end of the Pleistocene is regarded as an interglacial period (a warm spell with diminished glaciers) rather than the "end" of the Ice Age.

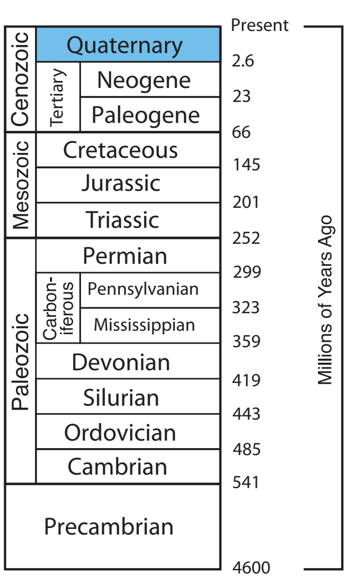
Figure 4.23 shows that after the global average temperature increased approximately 10,000 years ago, temperatures in the Northern Hemisphere began a slow, steady decline, which lasted for most of the last 7,500 years. These data (from ice cores from Greenland) support the hypothesis that, at least in the Northern Hemisphere, if natural climate variation were unhindered by humans, temperature would remain nearly constant, or would very slowly decline over the next few thousand years.

3.5.6 The Last 1000 Years

As scientists look closer to the present, climate resolution continues to improve. As indicated by ice-core temperature proxies, the last 1,000 years show very little significant global average temperature change prior to the 20th century (see *Figure 4.24*). A careful eye might notice, however, that the period from year 1000 to 1300 CE is slightly warmer and more "noisy" than from 1300 to present. This represents a time called the **Medieval Warm Period**, during which the Viking people inhabited Greenland. In the centuries following (from 1400 to 1800 CE), in contrast, fisherman living in nearby Iceland, where the weather is more temperate, were unable to leave their fishing ports for up to three months out of each year. Although the Medieval Warm Period resulted in only a small increase in global average temperature, it was substantial enough in that region to impact the lives of the people living there.

In the last 100 to 150 years, global average temperature has dramatically increased (*Figure 4.24*). This increase is a significant departure from the trend of the last

Find more on recent climate change in Chapter 5: Evidence for Recent Climate Change.



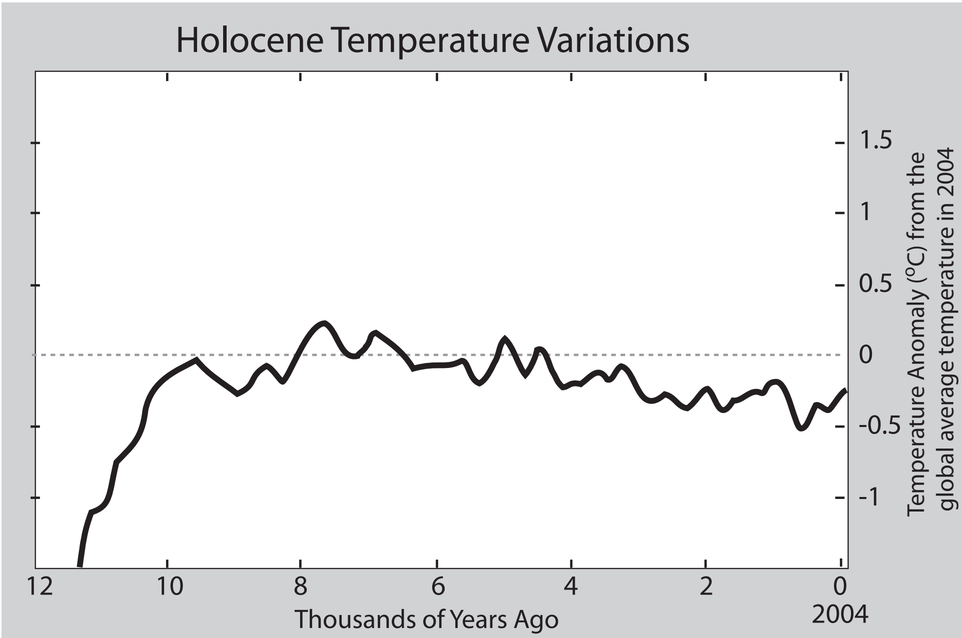


Figure 4.23: Holocene temperatures. Proxy temperature reconstruction from ice-core data after the end of the last glacial period, measured in °C difference from the global tempera¬ture average in 2004.

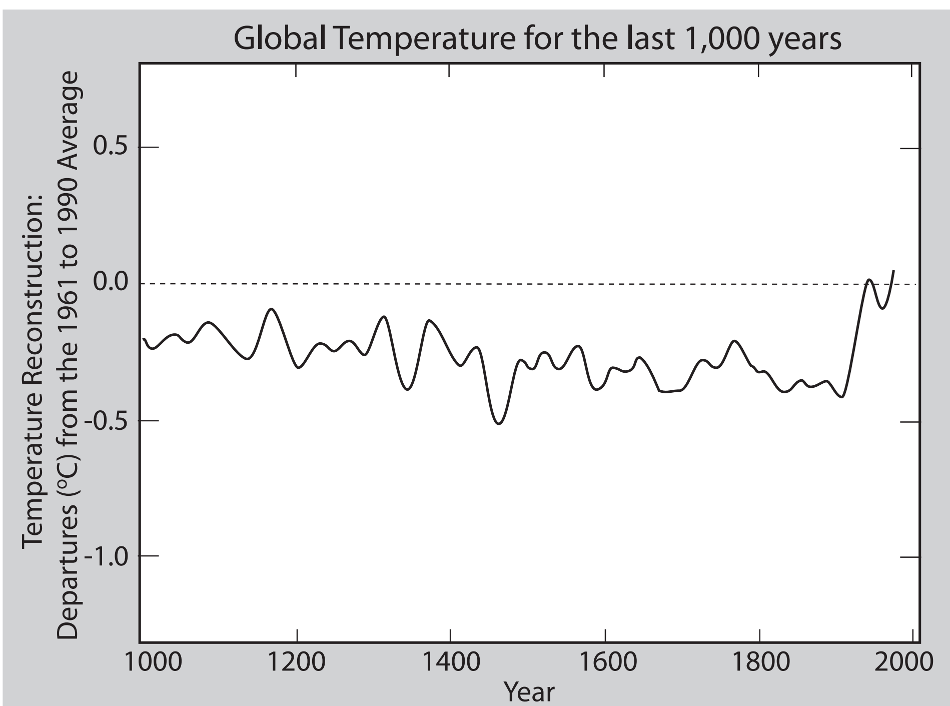
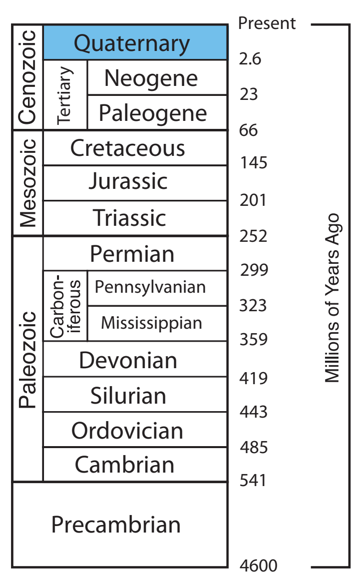


Figure 4.24: Global temperature for the last 1,000 years. Global temperature fluctuations in °C for the last 1,000 years relative to the global average temperature from 1961 to 1990, based on oxygen isotopes from ice cores.



Box 4.8: Using living organisms to determine past temperatures

Scientists can also use proxy records stored in living things and their ancient ancestors to recreate climates of the recent past with an amaz¬ing degree of precision.

Dendrochronology—the study of climate change as recorded by tree ring growth—is an excellent example of how climate researchers get information about climates of the relatively recent past. Trees can live for hundreds of years and in some extraordinary cases, like giant se-quoias and bristlecone pines, thousands of years. Each year, a tree adds a layer of growth between the older wood and its bark. This layer, or "ring" as seen in cross section, varies in thickness. A wide tree ring records a good growing season, usually moister and/or warmer, where-as a narrow ring records a poor growing season, usually drier and/or cooler. Especially in environments near the edge of a tree's comfort¬able living range, such as near the treeline on a mountain, where differences in weather make a large difference in growth rates, these data can provide highly reliable records of climate patterns.

To know more about climate over an even longer period of time, scientists look at dead trees that are still well preserved. They correlate the dead tree rings with the rings of a living tree whose age is known, looking for overlapping patterns, and can thereby get a longer record of climate through time. An amazing example is seen in the tree-ring chronologies established by looking at bristlecone pines from the southwestern U.S. (*Figure A*). Not only are these trees the longest-lived trees on Earth, they also live in a place where, even when they do die, they remain well preserved for hundreds or thousands of years. By comparing rings of living to dead bristlecone pines, scientists have established a tree ring record of climate in the Western U.S. for the past 9,000 years.

Figure A: Bristlecone pines, like this Pinus longaeva*, in Ancient Bristlecone Pine Forest, White Mountains, California, are among the longest-lived organisms on Earth. Their tree rings have provided scientists with climate data for the past 9,000 years.*

Box 4.8: Continued

There are, however, limitations to dendrochronology. Trees in the temperate zone only record the growing season, so the winter season is not usually visible in their wood. Trees in tropical regions grow year-round and therefore show no obvious annual growth rings, so climate data from equatorial areas is more difficult to obtain. And the variable most responsible for variable tree growth rate (temperature, moisture, or another factor) is not always unambiguous, thus research requires large data sets to overcome the uncertainty of individual data points.

Living Coral as a Climate Proxy

Because they build their own calcium carbonate ($CaCO_3$) skeletons, corals (*Figure B*) keep a record of climate in a way very similar to trees – by periodic rings of growth in the skeleton. Thicker rings represent better conditions for the coral, whereas thinner rings represent poor conditions. The coral colony grows both in winter and in summer, but the density of the skeleton is quite different due to seasonal changes in ocean temperature, the availability of nutrients, and differences in light. Additionally, the coral rings contain carbon and oxygen isotopes that indicate environmental conditions at the time that that part of the skeleton was secreted.

Figure B: A living colony of pillar coral at Discovery Bay, Jamaica. Such colonies can be hundreds of years old and the characteristics of older layers can provide proxy evidence of past environmental conditions.

800 to 900 years, an interval without major anthropogenic influence. In fact, *the warming seen in the last 50 years may be more pronounced than at any time in at least the last 65 million years*.

4. Climate Analogs and Models

We cannot do global experiments on climate processes, but we can observe what sort of processes have occurred during a wide range of Earth system changes in the past. More recent analogs are strong because the starting conditions (such as continental positions and ecosystems) are more similar to today, and also because recent analogs often provide more and better preserved data. On the other hand, some scenarios that are less perfect analogs provide scenarios that may not be present in more recent records, such as rapid increases in methane (Paleocene-Eocene) and climate-related mass extinctions (e.g., end-Permian). Because each analog has different pros and cons, paleoclimatologists collectively study a variety of past scenarios toward better understanding of current climate change.

We can also use paleoclimate data to test computer climate models by observing how well the models reproduce actual past climate changes, at either ecological or geological time scales, in order to better gauge how accurately they may estimate future changes. Some models are focused not specifically on predicting future climates, but on reproducing past climates well enough for us to understand how the evolving climate system works. These models are important because they could enable us to consider what may happen to future climates, even if past climates are not good analogs in some respects.

Resources

Books

A technical yet relatively readable book put together by a committee for the National Research Council on the potential of the deep-time geological record to inform us about the dynamics of the global climate system: Understanding Earth's Deep Past Lessons for our Climate Future, 2011, National Academies Press, Washington, DC. Available as a free download from the National Academies Press: http://www.nap.edu/download.php?record_id=13111.

The standard college textbook on the history of climate is by William F. Ruddiman Ruddiman, 2014, Earth's Climate: Past and Future, 3rd edition, W. H. Freeman, New York, 445 pp. A comprehensive review of the research literature can be found in Paleoclimates: Understanding Climate Change Past and Present, by Thomas Cronin, 2009, Columbia University Press, New York, 448 pp.

Online Resources

The Howard Hughes Medical Institute (HHMI) devotes one module of its program Changing Planet Past, Present, and Future to “Paleoclimate: A History of Change”: http://www.hhmi.org/biointeractive/paleoclimate-history-change.

Paleomap Project: a set of detailed maps of the world showing the past positions of the continents and describing Earth’s past climates, going back to the Cambrian period: http://www.scotese.com/climate.htm.

The Paleontology Portal: North American fossil record and geologic and climate histories. Use the Exploring Time & Space section to investigate geological history state by state and period by period: http://paleoportal.org/.

Paleo Perspectives: a series of informational websites on paleontological perspective on modern issues, including “A Paleo Perspective on Global Warming”: https://www.ncdc.noaa.gov/data-access/paleoclimatology-data/perspectives.

Geologic Time

There are numerous resources on teaching geologic time, many of them focused on putting Earth history events along a timeline of a convenient scale. The Texas Memorial Museum at the University of Texas at Austin website has a pdf “Understanding Geologic Time” with timeline activities for middle school students: http://www.jsg.utexas.edu/glow/files/Understanding-Geologic-Time-6-8.pdf.

Climate and Earth History

Graphing Paleoclimate

This series of lesson plans is designed for 9th grade students to better understand the factors that drive climate change over geologic time: https://scied.ucar.edu/graphing-paleoclimate.

Paleoclimates and Pollen

In this activity directed at 7th to 9th grade students, students learn about how fossil pollen could be used to understand ancient climates: https://scied.ucar.edu/activity/paleoclimates-and-pollen.

Anthropocene

The Anthropocene is the interval since humans became a major geologic force and the term is under study for formal recognition by the International Stratigraphic Commission. Educational resources for discussing with students the nature of human impacts, including climate change, can be found at the HHMI website: http://www.hhmi.org/biointeractive/anthropocene-human-impact-environment.

Critical Zone Science

The critical zone extends from the bottom of the water table to the tops of vegetation, and critical zone science the interactions among air, water, life and rock, over time scales of seconds to geologic time. The CZ and how it is studied is well-suited to addressing The Next Generation Science Standards (NGSS). The Critical Zone Observatory website contains K-16 resources for integrating the critical zone into science education. http://criticalzone.org/national/education-outreach/k-12-education-1national/.