CHAPTER 12

Long-Term Climate Regulation

Key Questions

- Why was Earth's climate warm despite reduced solar luminosity in the distant past?
- Was Earth's surface ever totally frozen?
- Has Earth's climate generally been warmer or colder than today's climate?
- Why was the climate warm during the time of the dinosaurs, and why has it cooled over the past few tens of millions of years?

Chapter Overview

Solar evolution models predict that the Sun was about 30% dimmer when it first formed and that its luminosity has increased more or less linearly since then. Nevertheless, Earth appears to have had liquid water at its surface for as far back as the geologic record extends, roughly 4.2 b.y. Warm temperatures on the early Earth were maintained by a combination of powerful greenhouse gases, especially CO_2 and CH_4. CH_4 concentrations fell abruptly around 2.4 b.y. ago when atmospheric O_2 levels rose, throwing Earth into its first major glaciation. Indeed, this glaciation appears to have been the first of three or more "Snowball Earth" episodes, during which times Earth's surface was entirely covered with ice and snow. Earth recovered from this climatic catastrophe because volcanic CO_2 accumulated in the atmosphere, thereby increasing the greenhouse effect and eventually melting the ice. As the Sun gradually brightened, atmospheric CO_2 concentrations diminished, maintaining the climate within the limits favorable to life. This decrease in CO_2 was not accidental; rather, it was a natural consequence of a negative feedback in the carbonate–silicate cycle. Over the past few hundred million years, Earth's climate has fluctuated between warm and cold conditions, primarily as a result of changes in atmospheric CO_2 induced by plate tectonics and the carbonate–silicate cycle.

INTRODUCTION

Earth has been in existence for billions of years and has been inhabited by organisms for most of that time. All known organisms require water during at least part of their life cycles (although some can do without water for extended periods). This implies that Earth's surface temperature has remained warm enough to support liquid water throughout this entire time. But we have already seen that the Sun was approximately 30% less bright early in the solar system's history. Why did the climate remain relatively warm despite such a large change in solar heating?

Based on what we learned in Chapter 3, the solution to the "faint young Sun paradox" probably involves either a stronger atmospheric greenhouse effect or a lower planetary albedo, which is what is needed to balance the planetary energy budget. When we last looked at this problem, we had no good reason to believe that either of these factors should have changed in the needed direction. In Chapter 8, however, we studied another component of the Earth system, the carbon cycle, in some detail. In particular, we found that a strong negative feedback exists between atmospheric CO_2 concentrations and the rate of silicate weathering. This feedback has a tendency to stabilize climate over long time scales because it causes atmospheric CO_2 levels to increase when the climate gets too cold or to decrease when it becomes too warm.

This negative feedback in the carbonate–silicate cycle is a plausible solution to the faint young Sun paradox. But it cannot explain all the details of Earth's climate evolution. As we saw in the last chapter, there are good reasons to believe that methane was much more abundant prior to the rise of atmospheric O_2. Methane is a good greenhouse gas; hence, it may well have played a role in keeping the early Earth warm. Furthermore, if we examine Earth's climate history more closely, we find that Earth's surface temperature has experienced substantial swings: there have been periods (like the present) when the polar regions have been covered with ice, and there have been other, longer periods when polar ice appears to have been completely absent. There have even been periods, like the Neoproterozoic era (about 700 million years ago), when continental ice sheets may have existed in the tropics! In this chapter, we examine the question of just how stable Earth's climate has been, and we look at how plate tectonics and biological evolution may have caused large climatic shifts.

THE FAINT YOUNG SUN PARADOX REVISITED

Let us return now to the problem that we considered briefly at the end of Chapter 1: the faint young Sun paradox. We have already mentioned that the Sun gets brighter as it ages. The luminosity increase is a direct consequence of the density change caused by the conversion of hydrogen into helium. As such, it is considered to be a *robust* prediction of solar evolution models, meaning that it does not depend sensitively on the details of the model. Thus, even if physicists are wrong about precisely how fusion occurs in the Sun's core, the faint young Sun problem is not likely to go away.

How Well Do We Understand Solar Evolution?

Recently, our faith in solar evolution models was increased by an important new observation. Physicists had been bothered for years about the apparent underabundance of neutrinos emitted by the Sun. **Neutrinos** are nearly massless particles emitted during nuclear reactions. The word "nearly" here is important. Until just recently, it was not known whether they had any mass at all. Like photons, they could have been totally massless. However, researchers at two new underground neutrino detectors, one in Sudbury, Canada, and another in northern Japan, have shown that this cannot be true. Neutrinos come in three different "flavors," only one of which was able to be measured by earlier detectors. These detectors measured only about one-third as many neutrinos as they should have, based on our theories of what is happening in the solar interior. However, if neutrinos have mass, then they can convert to the other two forms of neutrinos on their way from the Sun's core to Earth. The Super-Kamiokande detector in Japan measured all forms of neutrinos, whereas the Sudbury detector measured only one type. By comparing the numbers in the two experiments, the researchers were able to show that the total number of neutrinos emitted by the Sun agrees with the number that is predicted theoretically. Thus, solar physicists have renewed faith that they understand how the Sun produces energy.

There is one way in which the faint young Sun paradox could be avoided or, at least, modified. If the Sun was a few percent more massive in the past, it would have been brighter than it is now. But this would imply that it must have lost large quantities of material during its lifetime. The Sun does lose mass by way of the **solar wind**, an outflow of charged particles from the Sun's **corona** (the hot, outermost layer that is visible during a solar eclipse; see Figure 12-1). But the solar wind mass flux is about 10,000 times too small to account for a 1% change in the Sun's mass over geologic time. Rapid mass loss does occur in young stars that are spinning rapidly. The rapid rotation, combined with the presence of the star's magnetic field, heat up the star's corona, and this drives a much stronger stellar wind. However, observations of nearby young stars by Brian Wood and his collaborators at the University of

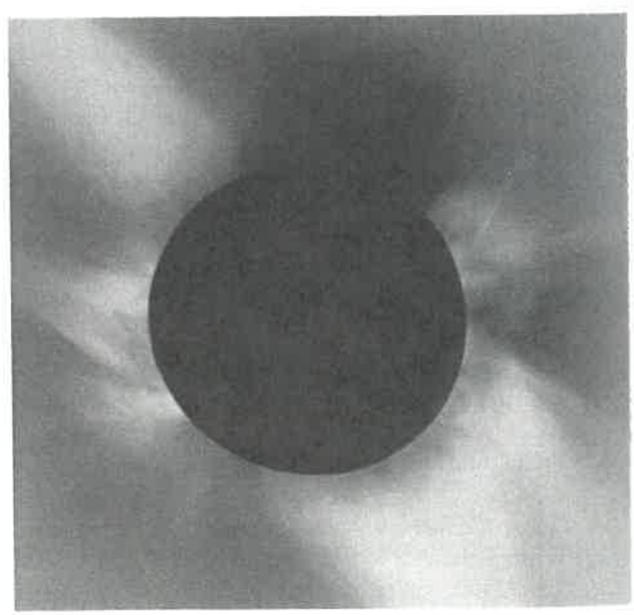


FIGURE 12-1 The solar corona, as revealed during a solar eclipse. (*Source:* UCAR/NCAR/High Altitude Observatory.)

Colorado have shown that these strong stellar winds persist for less than 200 m.y. After that, the stars spin more slowly because they have been slowed by the drag from their magnetic fields. Once the star's mass has stabilized, its luminosity rapidly approaches that predicted by standard stellar evolution models, implying that stars like our Sun are significantly less bright than they will be when they reach the Sun's age. Thus, the faint young Sun paradox appears to require an Earthbound explanation.

Defining the Faint Young Sun Problem Mathematically

Let us see if we are any closer to resolving this paradox now that we have examined the Earth system in more detail. If you worked out "Critical-Thinking" Problem 5 in Chapter 3, you should have determined that a 30% reduction in solar luminosity would have led to a 22° decrease in mean surface temperature, T_S, if Earth's albedo and greenhouse effect remained unchanged. As T_S is 15°C today, this would make the surface temperature at 4.5 b.y. ago equal to about −7°C, well below the freezing point of water. The actual surface temperature could have been even colder because of the positive feedback provided by water vapor and ice cover. A climate model calculation that includes the water vapor feedback (by assuming fixed relative humidity) is shown in Figure 12-2. The solar luminosity curve is the same as the one shown in Chapter 1. The lower dashed curve represents the effective radiating temperature, T_e, calculated using the principle of planetary energy balance from Chapter 3. The upper dashed curve is the mean surface temperature calculated by the climate model. The shaded region between the two curves represents the greenhouse effect, ΔT_g; this increases with time

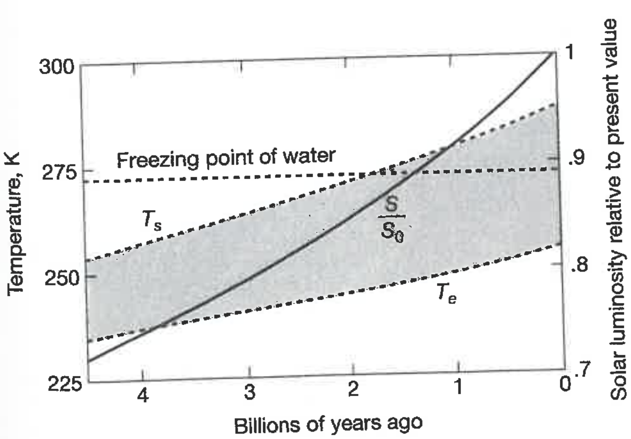


FIGURE 12-2 The faint young Sun paradox. The scale on the right applies to the solar luminosity curve, labeled S/S_0; the scale on the left applies to temperature curves. The shaded area represents the magnitude of the atmospheric greenhouse effect. (*Source:* From J. F. Kasting et al., How Climate Evolved on the Terrestrial Planets. *Scientific American* 256[2]:90–97, 1988. Used with permission, © George V. Kelvin/*Scientific American*.)

because of the water vapor feedback. A constant atmospheric CO_2 concentration of 340 ppm and a constant surface albedo are assumed.

Under these assumptions, T_S falls below the freezing point of water prior to 1.9 b.y. ago and reaches a chilly 255 K (−18°C) at 4.6 b.y. ago. This prediction is at odds with geologic evidence discussed in the previous two chapters, which shows that liquid water was present at 3.8 b.y. ago and that life has probably been present since that time as well.

Possible Solutions to the Problem

How can the faint young Sun problem be resolved? According to the planetary energy balance equation, three types of solutions exist. Either the planetary albedo must have been lower in the past, the greenhouse effect must have been larger, or additional heat sources besides the Sun must have been present.

One additional heat source that is sometimes suggested for warming the early Earth is **geothermal heat** from Earth's interior. As mentioned in Chapter 7, some of this heat is produced by the decay of radioactive elements in Earth's crust and mantle and some is left over from Earth's formation. Both heat sources should have been larger in the distant past, so it is not unreasonable to ask whether geothermal heat might have helped keep Earth warm.

The problem with this suggestion is that the geothermal heat flux is simply not big enough to supply the required energy. A deficit of 30% in the solar flux is equivalent to a loss of about 70 W/m^2 of heating, averaged over Earth's surface. The modern geothermal heat flux is only about 0.09 W/m^2. Theoretical models of Earth's interior evolution suggest that the geothermal heat flux at 4 b.y. ago was higher than the present value by a factor of 3 or 4, so the available heat flux should have been approximately 0.3 W/m^2. The release of this much heat at the seafloor would have prevented the oceans from freezing to the bottom, but they should still have been covered with an ice layer several hundred meters thick if no additional surface warming was present. Little or no light would have penetrated such an ice layer, so this would be hard to reconcile with the evidence for ancient photosynthetic life. (Recall from Chapter 11 that stromatolites provide evidence that some type of photosynthetic bacteria were already extant by 3.5 b.y. ago.) The early Earth was not a global iceball.

It is also difficult to solve the faint young Sun problem by simply changing Earth's albedo. The current planetary albedo is about 0.3, so the factor $(1 - A)$ in the energy balance relation is equal to 0.7. If the solar luminosity, S, were 30% lower, A would have to be near zero to keep T_e constant. It is difficult to imagine how a water-covered planet could have an albedo near zero, because clouds, which are highly reflective, would almost certainly have been present. If the surface was actually as cold as predicted by Figure 12-2, sea ice would likely have been present also, even if continents and continental ice sheets were not. So, although Earth's albedo may have changed with time, it is unlikely that this by itself could have kept the planet warm.

The most likely solution to the faint young Sun problem is that Earth's greenhouse effect was larger in the past. If so, though, which greenhouse gases would have been more abundant? As we have seen previously, water vapor is the strongest greenhouse gas in the modern atmosphere. However, water vapor cannot solve the faint young Sun problem by itself because it is close to saturation and, hence, acts as a feedback on climate rather than as a forcing. Sagan and Mullen, who pointed out the faint young Sun problem in the first place, suggested that ammonia, NH_3, might be the solution. Ammonia is a good absorber of infrared radiation and it is also a **reduced gas**, that is, one that combines with oxygen to form stable compounds. As discussed in the previous chapter, reduced gases should have been much more abundant prior to the rise of atmospheric oxygen around 2.4 b.y. ago. That is why Sagan and Mullen thought that ammonia might be a good candidate for warming early Earth. After their paper was published, however, other researchers demonstrated that ammonia would have been rapidly destroyed by ultraviolet radiation. Hence, it is unlikely to have been abundant enough to have provided the necessary warming.

A CO_2-Rich Early Atmosphere?

One greenhouse gas that could have kept early Earth warm is carbon dioxide. There are several reasons for suspecting that atmospheric CO_2 levels were originally much higher. As discussed in Chapter 10, smaller continents would have reduced the amount of land available on which to weather silicate rocks and to store carbonate rocks; thus, the CO_2 sink should have been smaller. Impact degassing of late-arriving planetesimals, along with enhanced volcanism on the hot, young Earth, would have created a larger CO_2 source. All these changes would have favored higher atmospheric CO_2 concentrations. On the other hand, weathering of the seafloor itself could have drawn down CO_2 levels by converting silicate minerals in carbonates and allowing these minerals to be subducted into the mantle. So, it is difficult to say whether atmospheric CO_2 concentrations would have been high or low during the first several hundred million years of Earth's history. There is also little geologic evidence with which to test one's theories. Climate during the earliest part of Earth's history remains largely a mystery.

Beginning about 3.8 b.y. ago, the geological record improves and one can draw inferences about climate with somewhat more confidence. As we have seen, the presence of liquid water and (possibly) of life suggests that additional greenhouse gases were present. CO_2 is among the most likely of these for the following reason. As discussed in Chapter 8, the carbonate–silicate cycle, which affects atmospheric CO_2 levels over long time scales, contains a strong negative feedback. If Earth's surface temperature were lower as a result of low solar luminosity, the rate of silicate weathering should have been slower, thereby lowering the CO_2 loss rate. CO_2 emitted from volcanos would have accumulated in the atmosphere until the global rate of silicate weathering balanced the volcanic outgassing rate. If Earth had ever become entirely ice-covered, silicate weathering would have ceased entirely and volcanic CO_2 should have accumulated in the atmosphere until the associated greenhouse effect became large enough to melt the ice. Thus, the Earth system has a natural way of recovering from global glaciation. Most of the time, the feedback is strong enough that global glaciation is avoided. When it does happen, however, the system apparently recovers in exactly the manner described.

How much atmospheric CO_2 would have been required to keep the early oceans from freezing? If CO_2 and H_2O were the only important greenhouse gases, then the minimum CO_2 level needed to compensate for a 30% reduction in solar luminosity is 0.3 bar—about 1,000 times the amount of CO_2 in the atmosphere today (Figure 12-3). Although this sounds like a lot, this amount of CO_2 is not large compared to the total amount of carbon available. The CO_2 stored in carbonate rocks today would produce a partial pressure of some 60 bars were it all present in the atmosphere. Only a small fraction (0.5%) of this CO_2 is needed to resolve the faint young Sun problem. Indeed, atmospheric CO_2 levels may initially have been much higher than this, as mentioned earlier. The upper limit of

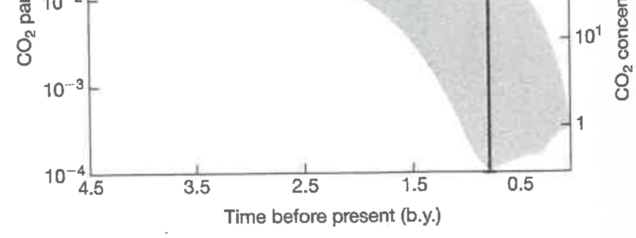


FIGURE 12-3 Atmospheric CO_2 concentrations needed to compensate for changing solar luminosity if CO_2 and H_2O were the only important greenhouse gases. The vertical bars at 2.5 and 0.65 b.y. ago show limits estimated from climate model calculations during glacial periods. (*Source:* Reprinted with permission from J. F. Kasting, *Science* 25:920–926. Copyright 1993 American Association for the Advancement of Science.)

10 bars shown in Figure 12-3 corresponds to the amount predicted on an ocean-covered early Earth. Climate model simulations show that such a CO_2 concentration would produce a global average temperature of 80–90°C. While we do not consider such a situation to be very likely, it is difficult to rule it out entirely. This would provide an alternative explanation for the prevalence of hyperthermophiles near the base of the evolutionary tree (Chapter 10, Figure 10-8).

Effect of Methane on Archean Climate

CO_2 was probably not the only greenhouse gas that affected Earth's early climate. We saw in the previous chapter that CH_4 could also have been relatively abundant prior to 2.3 b.y. ago, when atmospheric O_2 levels were low. This CH_4 could have come from a variety of biological and abiotic sources. Prior to life's origin, CH_4 could have been produced by impacts and by *serpentinization* of rocks on the seafloor. As mentioned in Chapter 10, serpentinization is a process by which various *serpentine* minerals are formed from reaction of water with iron- and magnesium-rich basalts. In the process, ferrous iron is oxidized to magnetite (Fe_3O_4) and water is reduced to molecular hydrogen (H_2). If CO_2 is present in the water, CH_4 is formed instead. These processes could have introduced modest amounts (10–100 ppm) of CH_4 into the prebiotic atmosphere. We saw earlier that this may have helped in the synthesis of the key biological precursor molecule, HCN.

Once life evolved, the source of methane to the atmosphere should have increased greatly. In Chapter 10 we argued that methanogenic bacteria were probably among the earliest organisms. These bacteria would have converted much of the H_2 in the atmosphere into CH_4 by way of the reaction $CO_2 + 4\ H_2 \rightarrow CH_4 + 2\ H_2O$. They could also have generated CH_4 from organic matter created by photosynthetic bacteria. Theoretical models suggest that atmospheric CH_4 concentrations of 1000 ppm or more are likely to have existed during the postbiological Archean and early Paleoproterozoic eras, 3.8–2.3 b.y. ago.

If atmospheric CH_4 was indeed present at these concentrations, it would have had a strong warming effect on global surface temperatures. The greenhouse effect of CH_4 would have been supplemented by warming from *ethane*, C_2H_6, produced from CH_4 photolysis. However, this greenhouse warming may have been offset to some extent by cooling caused by the presence of organic haze. Haze particles can produce an *anti-greenhouse effect*, as described further in the following text. The net result of these rather complex interactions is shown in Figure 12-4. The horizontal axis of the figure represents the atmospheric CO_2 partial pressure in bars (1 bar $\cong$ 1 atm). Recall that the partial pressure of a gas is just the pressure that it would exert if it was present by itself, with no other gases present. If the surface pressure is 1 bar, as it is in Figure 12-4, then a CO_2 partial pressure of 10^{-4} bar corresponds to a concentration of 100 ppm. The labels on the curves show the CH_4 *mixing ratio*. Mixing ratio is just fractional abundance, so a mixing ratio of 10^{-3} is equal to 1000 ppm. The solid curves show global average surface temperatures calculated with a one-dimensional radiative-convective model, or RCM. (Look back at Chapter 3 if you have forgotten what an RCM is.) The calculations were performed for an assumed solar luminosity of 80% of the present value, which is the value expected 2.8 b.y. ago.

What these curves demonstrate is the following: If CH_4 was *not* present in the atmosphere at that time, a CO_2

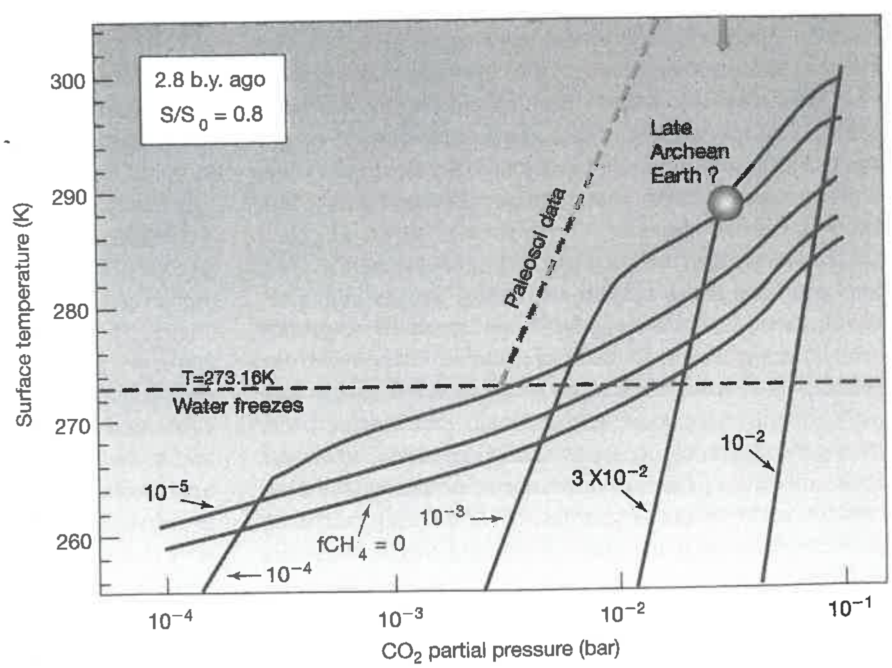


FIGURE 12-4 Average surface temperature as a function of atmospheric CO_2 and CH_4 concentrations. The total assumed surface pressure is 1 bar. The dashed curves show the freezing point of water and the published upper limit on Late Archean CO_2 derived from paleosols. The arrow at the top right shows a more generous upper limit on CO_2 derived from the paleosol data. (*Source:* J. Haqq-Misra et al., *Astrobiology*, v. 8, 2008, p. 1127.)

concentration of about 0.02 bar, or about 60 times the present level, would have been needed to keep Earth's surface from freezing at that time. If, however, CH_4 was present at a mixing ratio of 10^{-3}, as expected on the basis of biological considerations, then the surface temperature could have remained above freezing at a CO_2 concentration of 0.005 bar, or about 15 times present. Keeping the early Archean climate as warm as that of today (288 K) would require about 0.03 bar of CO_2, or about 100 times the present value. This point is marked by a star in Figure 12-4.

These attempts to model the climate of the early Earth are constrained to some extent by data. The nearly vertical dashed curve represents a published upper limit on atmospheric CO_2 levels derived from *paleosols* (ancient soils). Robert Rye and colleagues from Harvard University examined several paleosols of Late Archean age and noticed that none of them contained the mineral *siderite* ($FeCO_3$). Siderite is formed from the reaction between ferrous iron and carbonate ion ($CO_3^=$). Their analysis indicates that siderite should have formed in these soils if atmospheric CO_2 concentrations were higher than those indicated by the dashed curve. (Their limit is temperature-dependent because the chemical reactions involved in siderite formation depend on temperature.) All of the solid curves fall to the right of this line; hence, none of these solutions is able to satisfy this constraint. However, if the paleosols being studied were formed in tropical regions where the surface temperature was higher than average, then the upper limit on CO_2 partial pressure increases to 0.03 bar (100 times present), in agreement with the CO_2 concentration estimated previously. Thus, the study by Rye and colleagues is consistent with an Archean atmosphere in which CH_4 was an important greenhouse gas.

A Pink Sky during the Archean?

As mentioned in the previous chapter, high CH_4 concentrations could have made the Archean atmosphere appear very different from today's atmosphere. Today, the skies are blue as a consequence of scattering of sunlight by gas molecules (predominantly N_2 and O_2). **Scattering** is when a photon interacts with a particle (or molecule) and is sent off in a different direction. Gas molecules are smaller than the wavelength of visible light, so their interaction with sunlight is termed **Rayleigh scattering**. In Rayleigh scattering, the shorter wavelengths of light are scattered preferentially compared to longer wavelengths. Thus, if you are looking away from the Sun, the blue rays are scattered into your line of vision more effectively than are the red ones, giving the sky its bluish appearance. Conversely, when one looks directly at the Sun near sunrise or sunset, the blue rays are scattered out of your line of sight during their long pathlength through the atmosphere, and the Sun appears reddish-orange.

In the Archean, the interaction of sunlight with the atmosphere may have been quite different. When CH_4 becomes about one-tenth as abundant as CO_2 in an atmosphere, photochemical models predict that it can **polymerize** to form particles of higher hydrocarbons. Higher hydrocarbons are molecules in which carbon atoms are attached to each other in long chains. Planetary scientists believe that polymerization of CH_4 accounts for the orangish haze in Titan's atmosphere (see Figure 11-4 in the previous chapter). The reason that Titan's atmosphere appears orange is that the particles are approximately the same size as the longer (red) wavelengths of solar radiation. Scattering by particles that are comparable in size to the wavelengths being scattered is called **Mie scattering**. Mie scattering also predominates in Mars's atmosphere because of the large amount of suspended dust. Thus, the martian atmosphere looks pinkish, particularly during or after a global dust storm.

Earth's Archean atmosphere may have looked pinkish as well if the methane greenhouse story is correct. Indeed, there is a positive feedback that may have pushed Earth's climate system into a state in which organic haze would have started to form. Most methanogens are either **thermophiles** or hyperthermophiles. Recall that hyperthermophiles are organisms whose optimum growth temperatures are 80°C or above. Thermophiles are organisms with optimum growth temperatures between 40° and 80°C. Laboratory culture experiments have shown that the methanogens with higher optimum growth temperatures also have faster growth rates and shorter doubling times. Hence, a positive feedback loop exists whereby higher surface temperatures should have favored faster-growing methanogens. This, in turn, would have led to increased methane production, a bigger greenhouse effect, and hence still higher surface temperatures.

Climate Regulation by the "Anti-Greenhouse Effect"

The previous discussion makes it appear as though the early Earth should have become hotter and hotter until temperatures became too high for even the hyperthermophilic methanogens. (The highest temperature at which life has been found today is 121°C. Such temperatures can be reached without having the water boil if the overlying pressure is greater than one bar.) Well before this happened, however, another phenomenon would have occurred. As the methane content of the atmosphere increased, organic haze should have started to form. Both methane and the haze that it produces are strong absorbers of visible radiation. Methane itself absorbs in the red part of the visible and in the near-infrared. (This is why Uranus and Neptune appear blue. The red wavelengths of the incident sunlight are absorbed by the 2% methane in their atmospheres, so only the blue light is reflected.) Absorption of sunlight by methane and organic haze could have produced an **anti-greenhouse effect**, as occurs on Titan today. Indeed, the term "anti-greenhouse" was coined by Christopher McKay of NASA Ames Research Center in a paper about Titan's climate. In the anti-greenhouse effect, sunlight is absorbed high in the atmosphere and is reradiated back to space as infrared energy without ever reaching the planet's surface. This cools the surface, which is why it is called the "anti-greenhouse" effect. On Earth, the organic haze layer could not have been too thick, or it would have cooled the surface below the freezing point of water. Once this happened, the methanogenic bacteria that were producing the CH_4 would have died off, and the haze layer would have thinned.

All of this suggests that Earth's climate at this time may have been regulated by a negative feedback loop between surface temperature, atmospheric CH_4 and CO_2, and the organic haze layer (Figure 12-5). Higher temperatures would have led to an increase in CH_4 (for reasons described above) and to a decrease in CO_2. The decrease in atmospheric CO_2 with surface temperature is because of the well-known feedback involving the silicate weathering rate, which we have just been studying. As CH_4 levels went up and CO_2 levels went down, methane would have begun to polymerize and an organic haze layer would have begun to form. As soon as it became thick enough to block out sunlight, however, the anti-greenhouse effect would have set in and the surface would have begun to cool. The overall effect of this negative feedback loop should have been to stabilize Earth's surface temperature somewhere above the freezing point of water. Whether this would have kept the Archean Earth hot, or merely warm, depends on a variety of factors (e.g., continental size) that are difficult to determine. So, this hypothesis does not tell us what the Archean climate was like, but it does suggest why it may have been stable for long periods of time.

We note parenthetically that the Archean climate stabilization mechanism that we have just outlined is quite "Gaian" in nature—much more so than the present climate system. If methanogens truly were keeping the early climate warm, and if they themselves depended on this warmth in order to flourish, then they were clearly modifying climate in such a way as to benefit themselves. Indeed, we can emphasize the Gaian nature of the Archean climate system by redrawing the feedback loops (Figure 12-5) in the form of a Daisyworld-type diagram (Figure 12-6). In this diagram, surface temperature has been plotted on the vertical axis, unlike in Daisyworld (Figure 2-10), where it was plotted on the horizontal axis. The solid curve in the figure shows the effect of surface temperature on atmospheric CH_4. Higher temperatures lead to increased CH_4 production and, hence, to increased atmospheric CH_4. The parabolic curve in the figure shows the effect of atmospheric CH_4 on surface temperature. At low CH_4 concentrations, increased CH_4 causes an increase in surface temperature because it increases the greenhouse effect. At higher CH_4 concentrations, however, increased CH_4 causes a *decrease* in surface temperature by increasing organic haze formation and creating an anti-greenhouse effect.

The possible solutions to this Daisyworld-like diagram of the Archean climate system are the points P_1 and P_2. Point P_1 is unstable, however, as those who have completed "Critical-Thinking" Problems 2 and 3 in Chapter 2 will quickly realize. Point P_2 is the stable equilibrium point in this model. Because P_2 lies on the right side of the diagram, this model predicts that the Archean climate system should have stabilized at a point in which a thin organic haze was present in the atmosphere. So, in this case, "Gaia" has some predictive power! But is there any way of knowing whether this prediction is correct? To answer that question, we must examine more carefully the long-term geologic record of Earth's climate. The next section summarizes what we know, and do not know, about this intriguing subject.

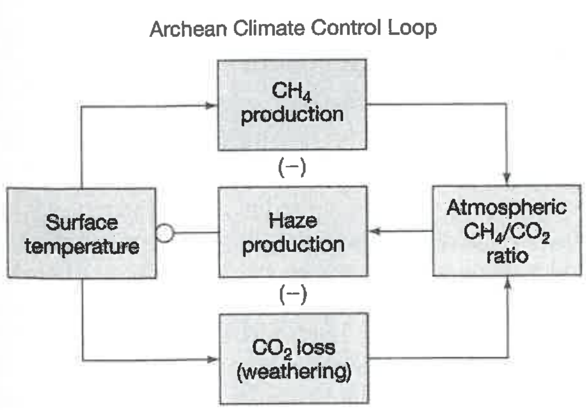


FIGURE 12-5 Feedback diagram showing a possible climate control mechanism that may have regulated Earth's surface temperature during the Archean era.

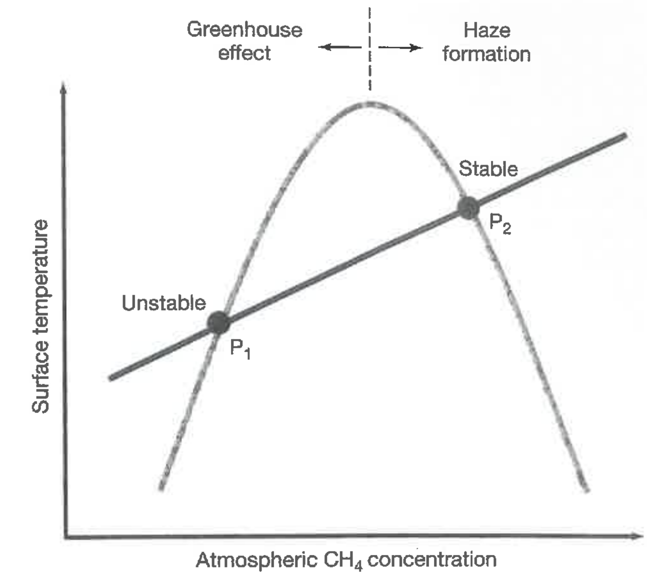


FIGURE 12-6 A "Daisyworld"-type diagram of the Late Archean climate system. See text for explanation. The diagram shows how the biota—methanogenic bacteria, in particular—may have helped stabilize the climate system at that time.

THE LONG-TERM CLIMATE RECORD

Our discussion to this point has been focused on the very early Earth and on processes that may have contributed to climate stabilization. After all, the most significant characteristic of Earth's climate on long time scales is that it has remained conducive to the presence of life for something close to 4 b.y. If one examines the geologic record in more detail, however, one finds that climate is anything but stable. A variety of different geologic indicators, discussed in the following text, show that **paleoclimate** (past climate) has actually varied in a complex manner, with long periods of warmth separated by shorter periods of intense cold. And there may have been several "Snowball Earth" episodes, when Earth's surface actually did freeze over entirely. This suggests that other factors that we have not yet talked about affect long-term climate as well. Determining what those other factors might be occupies the last two sections of this chapter. Before discussing them, let us examine some of the different types of geologic evidence that are used to study paleoclimate and see if we can piece them together to determine the broad outlines of Earth's climate history.

Geological Indicators of Paleoclimate

The types of geological indicators that are useful for determining paleoclimate depend on the time scale being considered. For recent periods of Earth history, it is possible to obtain a reasonably accurate estimate of ocean temperatures by measuring oxygen isotopes in carbonate sediments obtained from *deep-sea cores*. This technique will be described in Chapter 14, where it is used to identify the glacial–interglacial cycles of the past 3 m.y. This method works only for time periods more recent than about 200 m.y. because most of the seafloor older than this has been subducted.

For more ancient time periods, including the Precambrian, it is possible to look at oxygen isotopes in carbonate rocks and in cherts (SiO_2) preserved on the continents. If one takes these isotopic data at face value and analyzes them by the methods described in Chapter 14, one obtains a very curious result: Earth's mean temperature appears to have been about 70°C during the Archean eon and still 55–60°C throughout most of the Proterozoic! Some biologists are quite happy with this result, as it corresponds nicely with the biological evidence for a thermophilic or hyperthermophilic last common ancestor for extant life. These results, however, are difficult to understand, given the dimness of the young Sun. More importantly, they are in glaring contradiction with evidence (discussed below) for glaciation at several different times during the Precambrian. Hence, some researchers hypothesize that these isotopic data have been influenced by other processes, for example, by a change in the oxygen isotopic composition of seawater. The ongoing argument among scientists concerning these isotopic data is not one that is well suited for discussion in an introductory textbook. We mention it, though, simply to warn the reader that our understanding of the climate of the early Earth is still developing and that different researchers may have different opinions.

During the past 540 m.y. (the Phanerozoic era), we can learn about paleoclimate by examining the fossil record. Species of plants and animals that are known to live in certain climates can be used to estimate surface temperatures in the localities where their fossils are found. In doing so, we must account for the fact that the continents have drifted over Earth's surface as a consequence of plate tectonics. This technique is of limited use in the Precambrian era, however, because the single-celled organisms that were the only extant forms of life during most of this time could have survived under a wide range of climatic conditions.

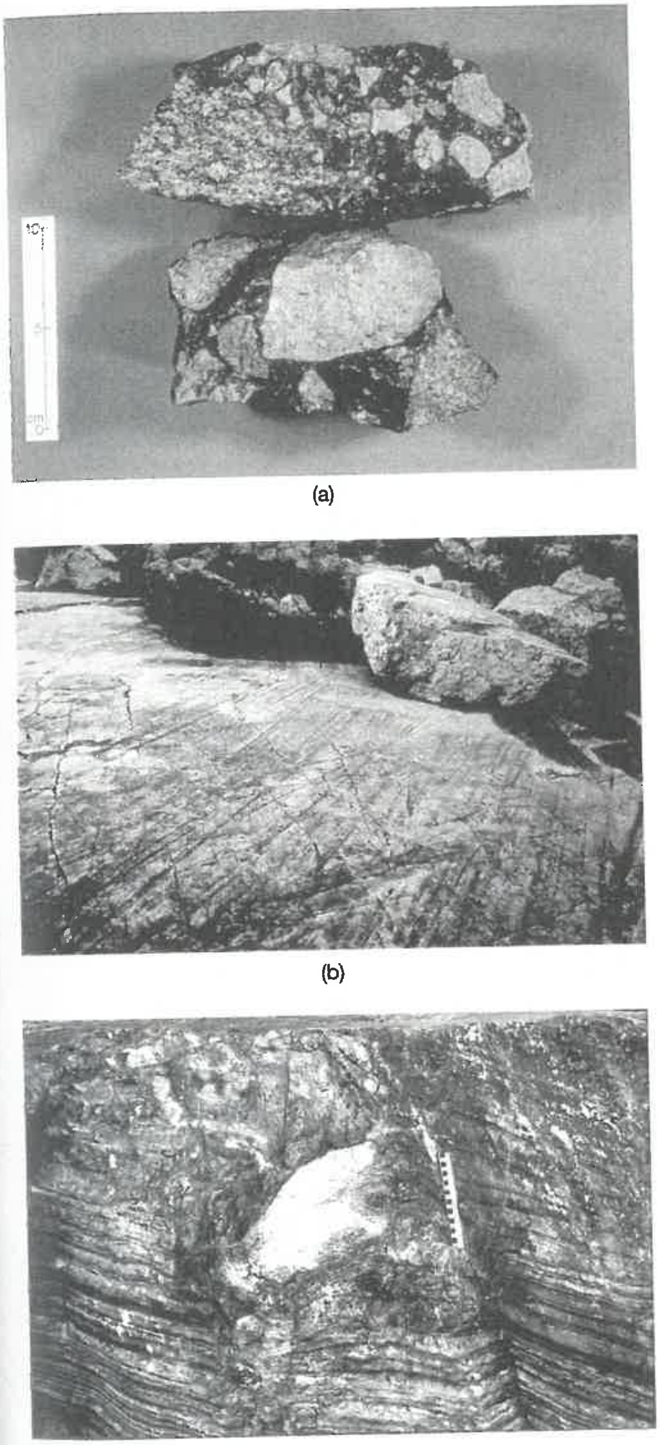
Evidence of Past Glaciations

The best available evidence for climate change on billion-year time scales comes from geologic deposits formed by glacial ice. Three such types of deposits are shown in Figure 12-7. **Tillites** (Figure 12-7a) are mixtures of cobbles, pebbles, sand, and mud that have been packed together to form rocks. They are formed from debris produced when glaciers grind up surface rocks. The debris is carried along by the glaciers as they move and is deposited in piles of rubble called **moraines** along the margins of the ice sheets. (Moraines mark the terminal points or flanks of the glaciers.) In association with tillites, geologists sometimes find rocks with long, parallel scratches, or **glacial striations**, formed when moving glaciers drag other rocks across their surfaces (Figure 12-7b).

A third signature of glaciation is **dropstones**, "misplaced" chunks of rock that occur in otherwise finely laminated marine sediments (Figure 12-7c). They form when rocks trapped in glacial ice are carried out to sea by icebergs, a process termed **ice-rafting**. When the iceberg melts, the trapped debris falls to the bottom of the ocean and becomes incorporated into sediments.

The Long-Term Glacial Record

Geologists who have studied the long-term glacial record have concluded that Earth's climate history has been marked by six main periods of glaciation (Figure 12-8). The first such period occurred at approximately 2.9 b.y. ago. This "Mid-Archean glaciation" is known from only two localities in South Africa and has not been studied as well as the later glacial intervals. The next glaciation, at 2.4 b.y. ago, is the time from which the tillite and dropstone pictured in Figure 12-7 derive. Because these deposits were first identified near the banks of Lake Huron in North America, this early cold period is sometimes called the "Huronian glaciation." The Huronian glaciation was



(a)

(b)

(c)

FIGURE 12-7 Geological indicators of glaciation: (a) a tillite from the 2.4 b.y.-old Gowganda formation in Canada; (b) glacial striations from the 0.65 b.y.-old Smalfjord tillite in Norway; (c) a dropstone from the Gowganda formation. (*Source:* (a), (b), and (c) J. William Schopf.)

followed by over 1 billion years of ice-free conditions during the Late Paleoproterozoic and Mesoproterozoic.

Why did Earth experience these major glaciations in the midst of what was otherwise an extended, ice-free stretch of its history during most of the Precambrian? Let's consider the Huronian glaciation first because that one is the easiest to understand. Indeed, the discussion we have just been through offers a convenient explanation. Suppose that methane *was* an important component of the atmospheric greenhouse during the Late Archean and early Proterozoic. The rise of atmospheric O_2 around 2.4 b.y. ago would have eliminated most of this methane, throwing climate into a temporary deep-freeze. And indeed, the geologic record is entirely consistent with this story. Figure 12-9 shows a stratigraphic section from the Huronian sequence of southern Canada. This particular section was deposited from 2.2–2.45 b.y. ago, based on radiometric age dating. The three bluish layers, G1, G2, and G3, represent glacial deposits (tillites), indicating that there were three separate episodes of glaciation. Beneath the lowermost glacial layer one finds rocks containing *detrital uraninite*. As discussed in the previous chapter, such deposits are indicative of a low-O_2 atmosphere. Above the uppermost glacial layer is the Lorraine *redbed* formation, which must have formed under a high-O_2 atmosphere. As pointed out over 40 years ago by Canadian geologist Stuart Roscoe, it appears as if the Huronian glaciation (or glaciations) is contemporaneous with the rise of O_2.

The Mid-Archean glaciation at 2.9 b.y. ago is not as easy to explain. One possibility, mentioned in the previous chapter (see the Box "A Closer Look: Mass-Independent Sulfur Isotope Ratios and What They Tell Us about the Rise of Atmospheric O_2" on p. 220 in Chapter 11), is that atmospheric O_2 increased just before this time, then went down again just afterward. This could have destroyed the methane greenhouse and thereby caused the glaciation, just as proposed for the Huronian. This might also explain the smaller $\Delta^{33}S$ values seen at this time (Figure 11-2). However, this explanation is inconsistent with other geologic evidence bearing on atmospheric O_2 levels. For example, detrital uraninite and pyrite are found in the Witwatersrand Basin of South Africa at essentially this time. As we saw in the previous chapter, such minerals are formed only under low-O_2 conditions. So, it seems more likely that the Mid-Archean glaciation was triggered by other factors.

Another idea that was also mentioned in Chapter 11 is that both the low $\Delta^{33}S$ values between 2.8 and 3.2 b.y. ago and the glaciation at 2.9 b.y. ago were caused by the presence of organic haze. Such haze could have blocked SO_2 photolysis, leading to low $\Delta^{33}S$, while at the same time creating an anti-greenhouse effect that might have triggered the glaciation. Although this idea is admittedly speculative, it would probably be easier to understand than would a transient rise of O_2. More detailed studies of sulfur

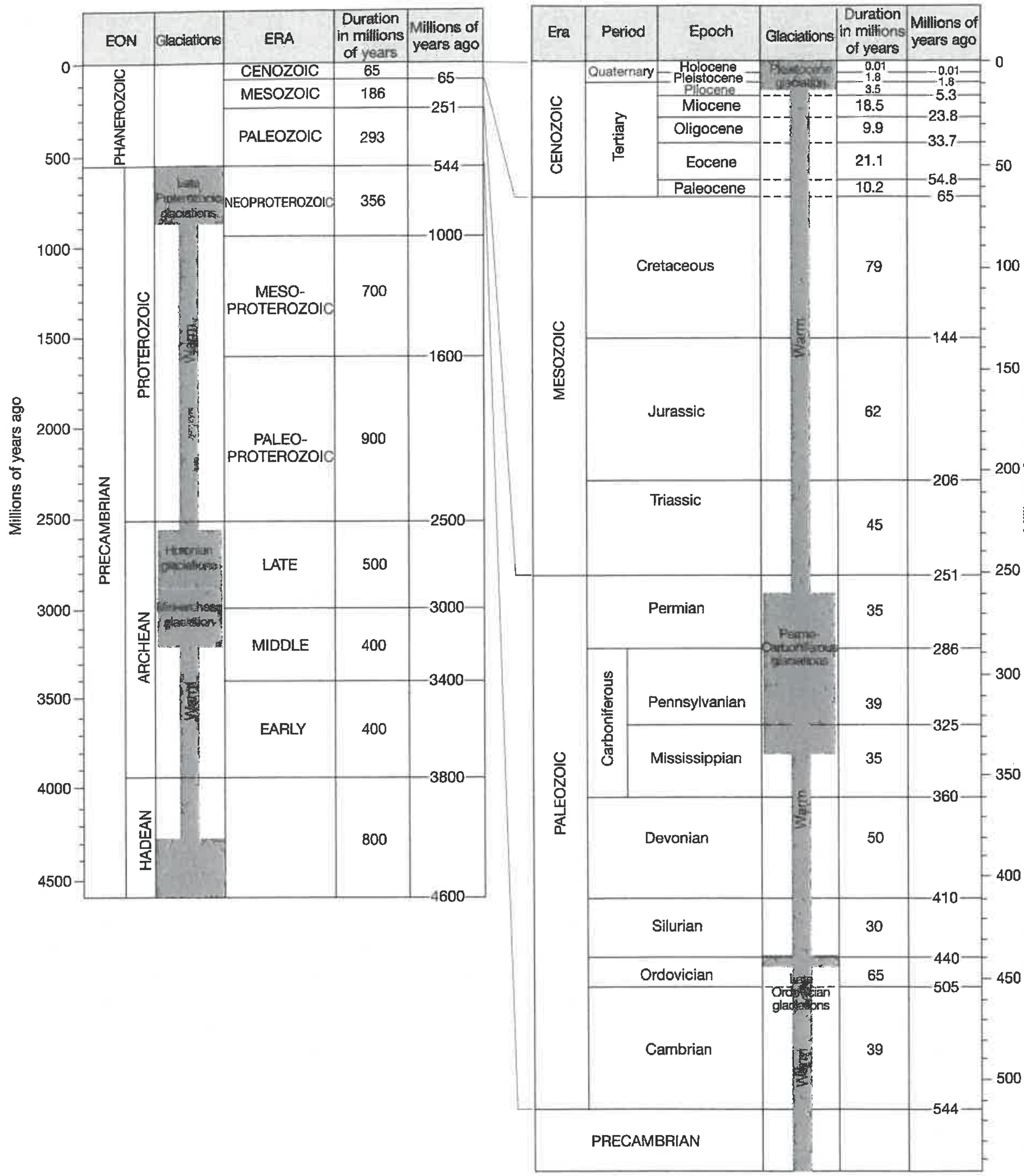


FIGURE 12-8 The major cold and warm periods during Earth's history. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

isotopes, looking at ^{36}S as well as ^{33}S, may help to show whether this hypothesis might be correct.

Let us move forward now to the early Proterozoic eon. Following the Huronian glaciations at 2.45–2.2 b.y. ago, the climate became warm again, as evidenced by the complete absence of evidence for ice. Why, one might ask, should the climate have warmed back up when the methane that was keeping it warm originally had largely disappeared? The carbonate–silicate cycle provides one possible explanation. CO_2 should have been outgassed from volcanos at about the same rate after the rise of O_2 as it had been before. Hence, it had to have been removed by

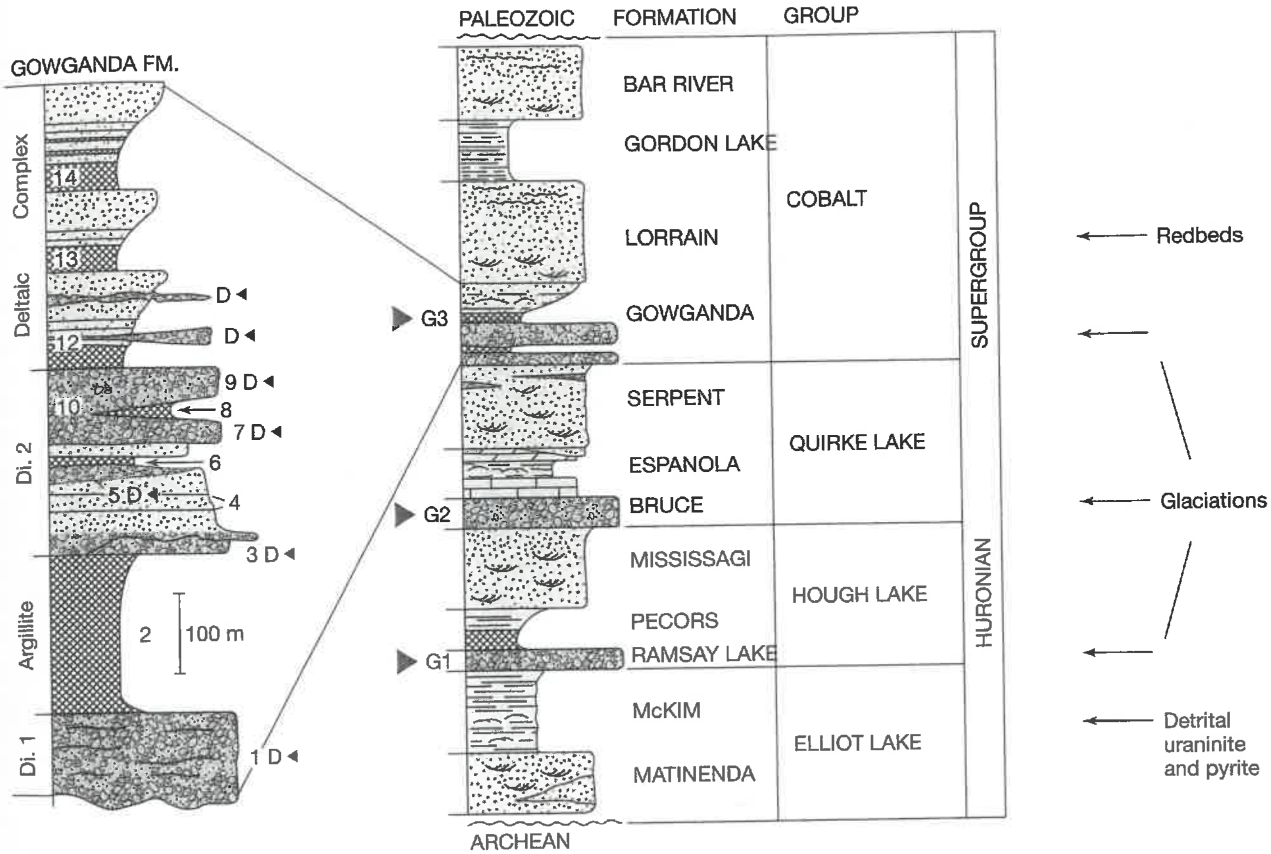


FIGURE 12-9 A stratigraphic section from the Huronian sequence in southern Canada. The bluish regions show the three glacial layers. (*Source:* G. M. Young, *Stratigraphy, Sedimentology, and Tectonic Setting of the Huronian Supergroup.* Field Trip B5 guidebook, joint meeting Geol. Assoc. Canada, Mineral. Assoc. Canada, Soc. Econ. Geol., Toronto, 1991.)

silicate weathering at that same rate. The silicate weathering rate is largely a function of surface temperature; thus, after the glaciation, the surface temperature had to recover to approximately its original value in order to keep the rate of silicate weathering the same as before. To accomplish this, atmospheric CO_2 could have jumped to a substantially higher concentration. This prediction remains speculative for the time being, but it might eventually be tested if more paleosol data become available to constrain past CO_2 concentrations.

There is another possibility for the extended warmth of the middle Proterozoic that climatologists have begun to think about only recently. It may be that after disappearing briefly just prior to the Huronian glaciation, atmospheric methane recovered and again reached levels much higher than today. Why, one might ask, should it have done so? A possible answer is that both atmospheric O_2 and dissolved oceanic sulfate levels may still have been much lower than today. This model for the Mid-Proterozoic has been championed by Donald Canfield and his colleagues from the University of Southern Denmark. Today, most of the organic matter that reaches the seafloor is recycled back to CO_2 either by aerobic bacteria that cause decay or by **sulfate-reducing bacteria**. (This is the process of bacterial sulfate reduction that was mentioned in the previous chapter.) If neither O_2 nor sulfate was present in appreciable concentrations, then much more organic matter may have decayed by the combined processes of fermentation and methanogenesis. (Methanogenesis is the biological production of methane.) These processes occur in modern marine sediments today, but only at depths below which the O_2 and sulfate are exhausted. So, today, little or no methane is released from marine sediments. But in the lower-O_2 Proterozoic, methane may have come out of marine sediments in large enough amounts to warm the atmosphere significantly and help keep Earth's surface ice-free.

Low-Latitude Glaciation: The Snowball Earth

Eventually the climate became cool once again. Indeed, the Late Proterozoic, between 0.75 and 0.6. b.y. ago, was so cold that it is considered a great mystery. Evidence for glaciation during this time interval is found on all seven present-day continents. The reconstruction shown in Figure 12-10 suggests that the continents were at that time grouped into two supercontinents, one of which was centered near the equator. Alternatively, the continents may

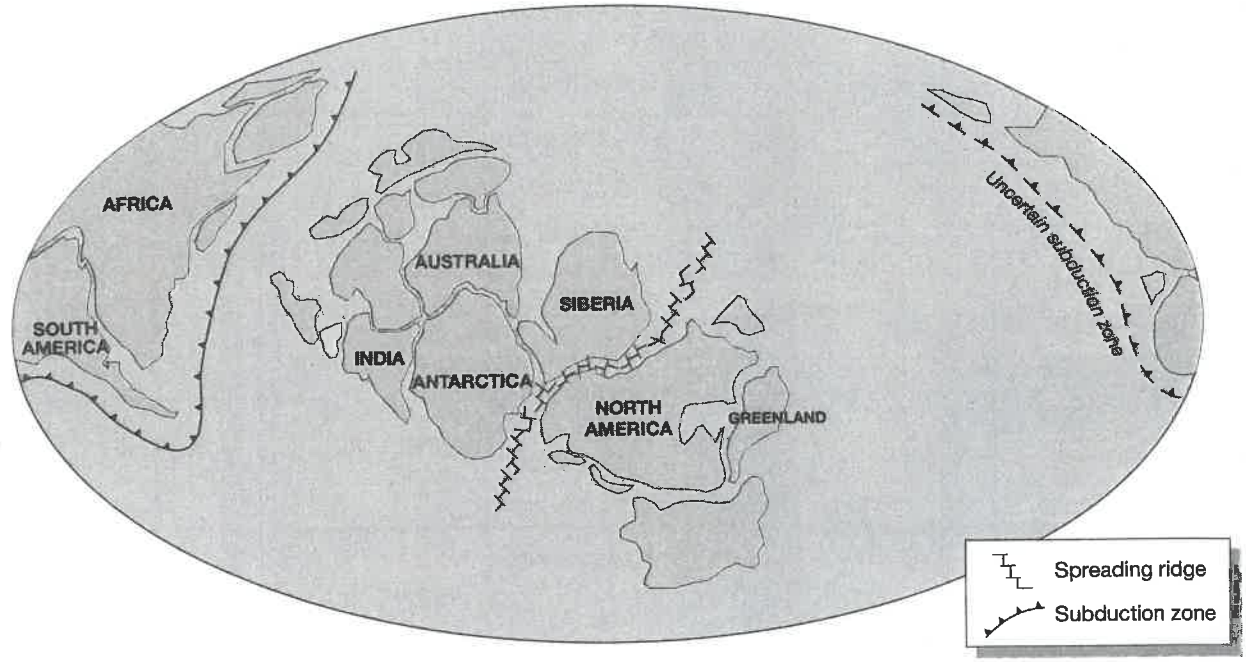


FIGURE 12-10 Possible continental reconstruction for the Neoproterozoic era. All the continents appear to have been glaciated at that time. (*Source:* J. L. Kirshvink, *The Proterozic Biosphere: A Multidisciplinary Study,* ed. J. W. Schopf and C. Klein, Cambridge: Cambridge University Press, 1992, Chapter 12.1.)

have been grouped into a single supercontinent centered on the equator but extending a significant distance to the north and south. In all continental reconstructions of this time period, one feature remains constant: The continent of Australia is situated at or near the equator. This is considered remarkable because the geologic evidence indicates that at this time Australia was glaciated from one end to another. The evidence includes tillites, glacial striations, and dropstones, the latter demonstrating that ice sheets extended right to the margins of the paleocontinent. Today, in contrast, there *is* glaciation in the tropics but it is confined to high mountain ranges like the Andes of South America. There, the glaciers never make it down below about 5 km elevation.

Why are geologists convinced that Australia was situated in the tropics at this time? Recall from Chapter 7 that past north–south continental positions are determined from paleomagnetic data. In Australia, glacial deposits are found mixed in with rocks in which the remnant magnetic field lines are parallel to the original bedding plane of the rock. If Earth's magnetic field was approximate to a dipole field, as it is today, this implies that these rocks formed near the equator. Many different geologists have looked at the paleomagnetic data from the Neoproterozoic rocks in Australia and have concluded that the evidence for low-latitude glaciation is real. The Huronian sequence described earlier also shows evidence for low-latitude glaciation; hence, it too may represent a "Snowball Earth" episode.

How could Earth have possibly gotten cold enough to glaciate the tropics? In one sense, it's not that difficult. After all, the entire first half of this chapter is devoted to the question of how Earth *avoided* global glaciation during the earlier parts of its history. If the Sun was less bright in the distant past, then the atmospheric greenhouse effect had to be larger, or else global glaciation would be expected. Conversely, if CO_2 and CH_4 levels were low at some time in the past, then the lower solar luminosity would ensure that Earth's climate was very cold.

Let us take the Neoproterozoic glaciations as an example because they are the ones that have been best studied. Fieldwork in Namibia, West Africa, by Paul Hoffman and his colleagues at Harvard University has shown that there were two main episodes of glaciation: one at ~730 m.y. ago and a second at ~610 m.y. ago. Consider the more recent of these two episodes. In the recent past, solar luminosity has been increasing at a rate of about 1% every hundred million years. Thus, the Sun should have been about 6% less bright at the time of the second glaciation. Under these reduced luminosity conditions, GCM climate modeling by William Hyde and colleagues from Texas A&M University indicates that CO_2 concentrations would have to have been more than twice the preindustrial level (i.e., 2×280 ppm) in order to avoid global glaciation. Other GCMs yield slightly different critical CO_2 levels, but all of them agree that there is some CO_2 level below which global ice cover cannot be avoided.

How exactly would a Snowball Earth glaciation proceed? To understand in detail, see the Box "A Closer Look: How Did Life Survive the Snowball Earth" on page 247. The general outline, however, is as follows. For one reason or another, atmospheric CO_2 concentrations were drawn down to relatively low values. (We will not consider the Huronian glaciation here. That one appears to have been caused by the rise of atmospheric oxygen.) Hoffman and his colleagues originally suggested that increased organic carbon burial on newly created continental shelves was the cause, but it appears likely that other factors were operative as well. Perhaps the most important of these is that a significant fraction of the continental area was situated in the tropics. This allowed silicate weathering to proceed even though Earth was growing colder and colder, so that atmospheric CO_2 could continue to be drawn down. This first step in initiating global glaciation is counterintuitive because the conventional wisdom regarding glaciations has been that they occur when a continent drifts near or over one of the poles. Global glaciations are different. They probably require continents at low latitudes. This may also explain why they have occurred only at certain times in Earth's history. When the continents are not concentrated at low latitudes, the silicate weathering feedback prevents global glaciations from occurring.

An alternative mechanism for triggering Snowball Earth, if one postulates that methane was still abundant during the Mid-Proterozoic, is that atmospheric O_2 may have increased near the end of this time period. Such an increase has been suggested for other reasons, as discussed in the previous chapter. (Increases in atmospheric O_2 are a possible trigger for the Cambrian explosion.) If atmospheric O_2 went up, the methane flux from marine sediments may have gone down, thereby reducing the methane greenhouse effect and making the climate colder. Such a change could conceivably have occurred faster than the carbonate–silicate cycle could regulate atmospheric CO_2. So, this again might have circumvented the normal processes that keep Earth's climate stable.

The rest of the sequence may have gone like this: As the Neoproterozoic climate became colder, for whatever reason, the polar ice sheets gradually crept down to lower latitudes. Once they reached approximately 30 degrees, however, something spectacular happened. All of a sudden, within a few decades perhaps, the oceans froze all the way down to the equator. The reason is that the positive feedback loop between ice albedo and surface temperature (Chapter 3, Figure 3-21) became so strong that it made the system unstable. This result can be demonstrated quantitatively using climate models. (See the Box "Thinking Quantitatively: Energy Balance Modeling of the Snowball Earth.") Once the ocean surface had frozen entirely, Earth's surface would have become extremely cold, −50°C or lower, because the albedo would have been very high (>0.6, as compared to 0.3 today) and most of the incident sunlight would have been reflected back to space. However, as soon as the surface froze, silicate weathering on the continents would have virtually ceased, and volcanic CO_2 would have begun to accumulate in the atmosphere. In approximately 10 m.y., given modern volcanic outgassing rates, the atmospheric CO_2 partial pressure would have reached 0.1 bar (300 times the current level) and, all of a sudden, the ice would have begun to melt. The positive ice–albedo feedback loop would now have worked in the other direction: increased melting would have led to decreased albedo, which would in turn have led to increased surface temperature and more melting. Models predict that the ice cover would have disappeared entirely within a few thousand years. At this point, Earth would have a dense, CO_2-rich atmosphere and a low albedo, and so it would have become very hot, with an average surface temperature as high as 50–60°C. Silicate weathering would now have proceeded rapidly, drawing down atmospheric CO_2 levels and eventually restoring the climate system to its original state.

Additional Geological Evidence for the Snowball Earth: BIFs and Cap Carbonates

To some, this story of an ice-covered Earth sounds simply too extraordinary to be true. Is there any additional evidence that such a sequence of events actually occurred? The answer is yes. A number of other observations are consistent with the Snowball Earth model. Of these, two features in particular stand out. The first is the reappearance of banded iron-formations (BIFs). Recall from the previous chapter that most BIFs were deposited prior to 1.8 b.y. ago and that their formation was linked to anoxic conditions in the deep ocean. Surprisingly, BIFs reappear briefly during the Neoproterozoic at exactly the time of the glaciations. This reappearance is explained by the Snowball Earth hypothesis because the global ice cover would have cut off the ocean from the large reservoir of atmospheric O_2. The dissolved O_2 in the oceans, or at least in some ocean basins, was used up by oxidation of organic matter in sediments. Ferrous iron emanating from the hydrothermal vents in the mid-ocean ridges accumulated in the ocean and was ultimately upwelled on continental shelves (after the ice had melted) and deposited as BIFs.

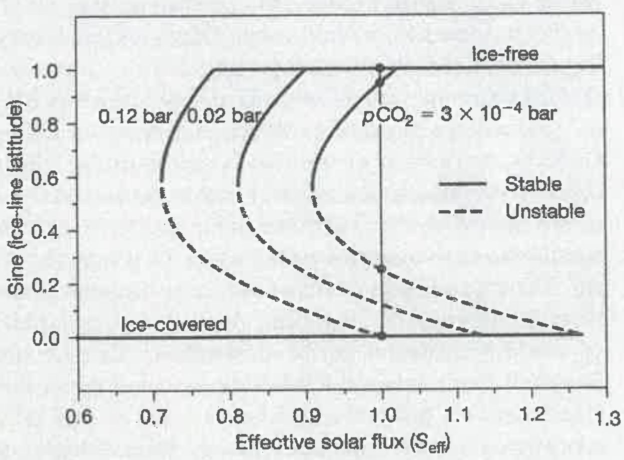
The other piece of evidence is even more telling. Directly above each of the two Neoproterozoic glacial deposits in Namibia is a layer of carbonate rock approximately 400 m thick. These carbonate deposits are often termed **cap carbonates** because they "cap" the glacial layers. The bottom parts of these caps are fine-grained and show evidence of having been deposited very rapidly, as would be expected in the immediate aftermath of a Snowball Earth episode. This close association between glacial deposits and carbonates has puzzled geologists for many years because glaciers typically form at high latitudes, where it is cold, whereas carbonates typically form at low latitudes where the surface ocean is warm and, hence, carbonate minerals are less soluble. The Snowball Earth hypothesis explains this association naturally; indeed, the model *predicts* that such carbonate deposits should have been formed. Thus, there is compelling evidence to indicate that the Snowball Earth model is correct.

THINKING QUANTITATIVELY

Energy Balance Modeling of the Snowball Earth

Although detailed modeling of the Snowball Earth climate requires a general circulation model, or GCM, much can be learned from simpler climate models. Indeed, much of the theoretical framework for understanding runaway glaciation was developed independently in the late 1960s by Soviet climatologist Michail Budyko and English climatologist William Sellers using what were later termed "energy-balance climate models," or EBMs. In a typical EBM, Earth's surface was divided into 18 different latitude bands, each 10 degrees wide. In their simplest form, these models calculated the annually averaged solar heating in each latitude band, along with the average outgoing infrared radiation flux. Heat transport between different latitude bands was parameterized as *diffusion*. Diffusion of heat is usually called *conduction*. We know, of course, that the atmosphere does not really transfer heat in this way. Rather, it does so by the complex system of winds and ocean currents described in Chapters 4 and 5. Budyko and Sellers did a clever thing, however: they adjusted their diffusion coefficients so that their models matched the observed equator-to-pole temperature gradient. Thus, their models were capable of reproducing the average latitudinal distribution of temperature and, most importantly, they could estimate the size of the polar ice caps. This allows such models to be used to study the phenomenon of runaway glaciation.

Results from a more up-to-date version of an EBM are shown in Box Figure 12-1. The figure is somewhat complicated, so let us go through it carefully. The horizontal axis is the effective solar flux (S_{eff}), that is, the solar flux divided by the modern value. Thus, $S_{eff} = 1$ corresponds to the modern solar constant. The vertical scale is the sine of



BOX FIGURE 12-1 Energy-balance climate calculations for the Snowball Earth model. Solid curves represent stable solutions; dashed curves represent unstable solutions. Dots show equilibirum solutions for today's solar flux.
(*Source:* K. Caldeira and J. F. Kasting, *Nature* 359, 1992, pp. 226–228.)

the ice-line latitude. This marks the extent of the polar ice caps. (The model is symmetric in each hemisphere.) Recall from trigonometry that sin 30° = 0.5, so a point halfway up the vertical scale corresponds to a latitude of 30°. Exactly half Earth's surface area is located poleward of 30°; the other half is located equatorward of this point.

The lines and curves in Box Figure 12-1 represent the ice-line extent as calculated by the EBM climate model. Solid curves represent stable solutions; dashed curves represent unstable solutions. The three different curves shown correspond to three different atmospheric CO_2 levels. The curve furthest to the right is for a CO_2 partial pressure of 3×10^{-4} bar, which corresponds to a CO_2 concentration of 300 ppm, close to today's value. The points where the curves (or lines) intersect a vertical line at $S_{eff} = 1$ represent stable climate solutions for the modern Earth. Surprisingly, there are three different, stable solutions. The one that actually corresponds to the modern climate is the "small ice cap" solution. The sine of the ice-line latitude is ~0.95, which puts the boundary of the polar ice cap at about 72°. But there are two other stable solutions as well: an ice-free solution (no polar cap) and an ice-covered solution. The ice-covered solution corresponds to the Snowball Earth. It is stable because the high albedo of the ice causes most of the incident sunlight to be reflected back to space.

The most interesting features of Box Figure 12-1, however, are the unstable solutions (dashed curves). As one can see, all solutions in which the ice line is equatorward of ~30° are unstable. If the polar ice caps ever reached this latitude, the ice–albedo feedback would have become completely unstoppable: increases in ice cover past this point would cause more sunlight to be reflected back to space, which would result in decreased surface temperatures and further increases in ice cover. Very quickly, within a few decades, the ocean surface would have frozen all the way down to the equator. This is thought to have been how the climate system became trapped in the Snowball Earth.

Box Figure 12-1 also shows how the system could have recovered from the Snowball Earth. Once Earth's surface was totally frozen, silicate weathering would have ceased and volcanic CO_2 would have accumulated in the atmosphere. One can see that when the CO_2 partial pressure reaches 0.12 bar (12,000 ppm), the unstable solution intersects the equator (sine ice-line latitude = 0) at $S_{eff} = 1$. Physically, this means that the ice-covered solution is no longer stable. Instead, the system would transition spontaneously (and rapidly) up to the ice-free solution. The climate would become extremely warm, 50–60°C, and would remain that way until silicate weathering was able to remove the excess CO_2 from the atmosphere. As described in the text, it looks as if this is exactly what happened during the Late Precambrian Snowball Earth episodes 610 m.y. and 730 m.y. ago.

A CLOSER LOOK

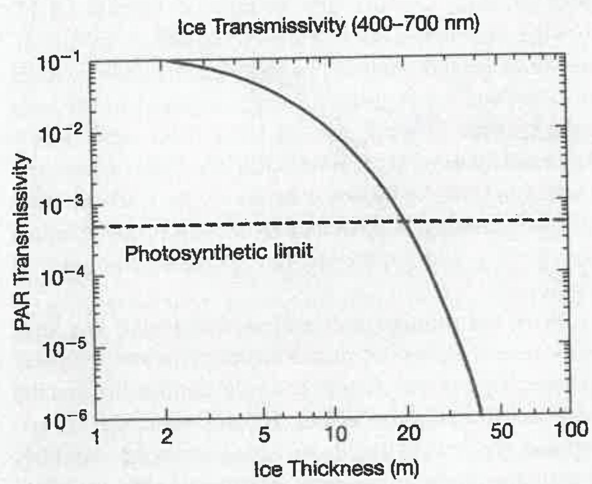
How Did Life Survive the Snowball Earth?

Perhaps the most interesting question regarding the Snowball Earth is: How did life manage to survive through it? One can estimate the thickness of the ice cover over the oceans by means of a fairly simple calculation. (See "Critical-Thinking" Problem 1 at the end of the chapter.) The thickness of the ice is limited by the geothermal heat flux that must be conducted upward through it. If you do the problem correctly, you should find that the sea ice during the Neoproterozoic glaciations was over a kilometer thick on average. This is far too thick to allow sunlight to penetrate. Very little sunlight makes it any deeper than 5 or 10 m even in very clear ice. And, yet, we are certain that photosynthetic life survived these two catastrophes, as well as the earlier Huronian glaciation, which may also have been a Snowball Earth episode. How can one resolve this apparent paradox?

One solution that has been suggested by William Hyde and his colleagues from Texas A&M University is that the tropical ocean may not have been entirely frozen. In their model (published in the journal *Nature* in May 2000), the tropical oceans remained ice-free, but at least some tropical continents were ice-covered. They suggested that the ice sheets formed at high elevations where the local temperatures were below freezing and that they flowed down to the continental margins before they melted. In this way, they could account for the occurrence of dropstones in Neoproterozoic marine sediments around Australia. Their model has been termed a "Slushball Earth" because the planet was never entirely ice-covered, if they are correct. However, their model has difficulty explaining the reoccurrence of BIFs and the presence of cap carbonates (see the next section); hence, it does not appear to us that their solution is the right one.

Another idea is that life survived in geothermally heated environments in which liquid water remained even as the rest of the surface froze. The hydrothermal vents of the mid-ocean ridges are one obvious refuge. As most mid-ocean ridge vents are over 2.5 km deep, they would have been well beneath the ice layer, and organisms living within them would have been essentially unaffected by a global glaciation. Such organisms are not photosynthetic, however. Continental geothermal areas such as Yellowstone Park in the United States are another possibility. However, the water in Yellowstone's geysers and hot pools ultimately derives from rainwater. During a Snowball Earth event, the hydrological cycle would have virtually shut down and such areas should have dried up. A few geothermal areas that are well connected to the ocean, the island of Iceland, for example, might have remained habitable for photosynthetic life. However, there are very few such areas in the world. If life had been restricted to only a few such locations, biologists believe that the Universal Tree of Life (Chapter 10, Figure 10-8) would show evidence for this. So, they think that photosynthetic life must have remained more widespread.

A third idea is that the ice may not have been as thick in all locations as simple models predict. Christopher McKay from NASA Ames Research Center near San Francisco has suggested that the ice could have been thin enough in the tropics to allow some sunlight to penetrate. Climate modeler David Pollard at Penn State has worked on this problem as well, along with one of the authors of this book (JK). McKay has spent several field seasons studying ice-covered lakes in the so-called dry valleys of Antarctica. The lakes there are covered with about 5 m of exceptionally clear ice, beneath which is found a thriving photosynthetic biota. The ice is clear because it forms very slowly and, hence, excludes air bubbles, which would otherwise give it a cloudy appearance. Box Figure 12-2 shows the fraction of light transmitted through such clear ice as a function of its thickness. If the ice in the tropics was as thick as predicted previously, then obviously very little light would make it through. But, if the ice was only a few meters thick, then as much as 10% of the incident sunlight might penetrate it. This energy would have to get out by conduction through the ice, just like the geothermal heat from Earth's interior. If you work "Critical-Thinking" Problem 2 at the end of the chapter, you should be able to show that a stable solution can exist with ice as thin as 2 m in the tropics. Approximately 10% of the incident sunlight could make it through this ice, according to Box Figure 12-2. This problem has now grown complicated, as one needs to also consider the possibility that thick ice near the poles—"sea glaciers," in the climate modeling lingo—would have flowed toward the equator. There is now a debate in the scientific literature as to whether or not this would preclude a "thin-ice" Snowball Earth. If such a solution is possible, however, this would readily explain how the photosynthetic algae made it through this catastrophe. So, perhaps this thin-ice model is indeed the solution to the Snowball Earth paradox.



BOX FIGURE 12-2 Visible light transmissivity (400–700 nm) versus depth for clear ice. (*Source:* C. P. McKay, *Geophysical Research Letters* 27, 2000, pp. 2153–2156.)

VARIATIONS IN ATMOSPHERIC CO_2 AND CLIMATE DURING THE PHANEROZOIC

Although the most spectacular extremes in climate appear to have occurred during the Precambrian, climate has varied over the past 542 m.y. as well. Three of the six glacial periods shown in Figure 12-8 have occurred during the Phanerozoic eon: a brief one during the Late Ordovician period (about 440 m.y. ago), a long series of glaciations near the boundary between the Permian and Carboniferous periods (about 280 m.y. ago), and the most recent episode, which is here called the Pleistocene glaciation. This name is something of a misnomer because Antarctica began to be glaciated some 15–30 million years ago, well before the Pleistocene epoch began, and because both Antarctica and Greenland continue to be ice-covered today. So, a better name for this glaciation (but one which would likely not be as widely understood) is the Late Cenozoic glaciation. To qualify as a long-term "glacial period," it is only necessary that ice be present over part of Earth's surface for an extended period of time. This is different from common usage, whereby the term "glacial period" refers to one of several episodes of maximum ice extent during the last 3 m.y. These shorter-time-scale climate fluctuations, which will be discussed in Chapter 14, can be thought of as modulations of an overall climate that is basically "glacial."

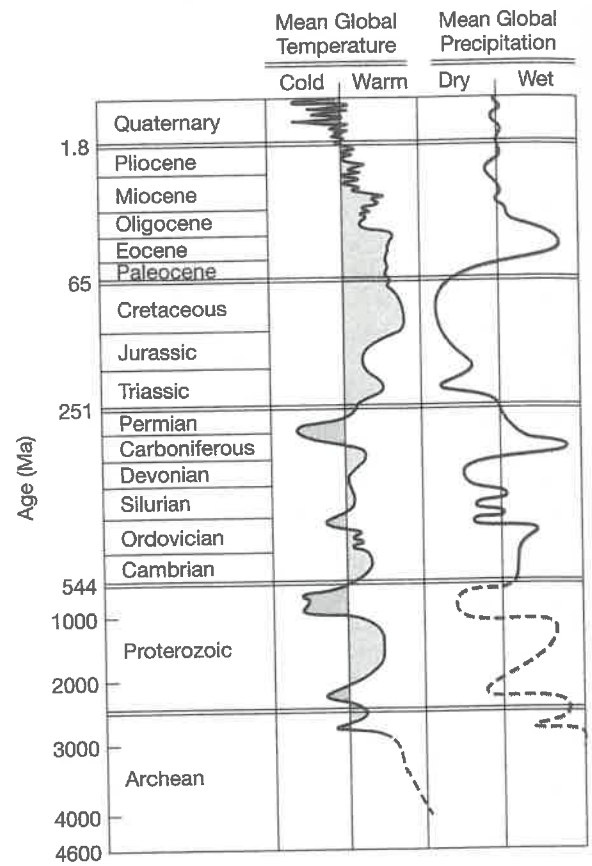


FIGURE 12-11 Estimated change in surface temperature during the Phanerozoic eon. (*Source:* From K. C. Condie and R. E. Sloan, *Origin and Evolution of Earth: Principles of Historical Geology*, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

CO_2 and Climate during the Paleozoic Era

The first part of the Phanerozoic eon, which extended from about 542 m.y. ago until 251 m.y. ago, is called the Paleozoic era. From this time on, considerable information is available to tell us about climate because the fossil record is much more detailed. By examining the types of plants and animals that existed and determining where they lived, scientists can deduce not only whether the climate was warm or cold, but also how wet it was and by how much it varied between the pole and equator. The general results of such studies are shown in Figure 12-11. Following the intense cold of the Neoproterozoic, the climate warmed. The Cambrian period was decidedly warmer than today, and the ensuing three geological periods were mostly ice-free as well, except for a brief (and poorly understood) spike of glaciation during the Late Ordovician. Climate cooled markedly, however, during the Carboniferous period, culminating in a series of glaciations that spanned almost 80 m.y. and are termed the "Permo-Carboniferous glaciations."

Why did climate get cold at this time? We have already seen that, on long time scales, climate is largely determined by a trade-off between solar luminosity and the atmospheric greenhouse effect. By this time, the atmosphere was well oxygenated, so we can assume that CH_4 concentrations were modest and that the greenhouse effect was mostly attributable to CO_2 and H_2O. Thus, the cooling in Earth's climate was probably caused by a decrease in atmospheric CO_2 levels.

Why should CO_2 levels have declined during the Late Paleozoic? On long time scales, atmospheric CO_2 is largely controlled by the carbonate–silicate cycle, but the organic carbon cycle cannot necessarily be neglected. Consider the carbonate–silicate cycle first. How might that have changed? As we saw in the previous section, the rate of silicate weathering is enhanced when continents move toward the equator. So, perhaps changes in continental positions were once again the key. But, biological innovations could have been important as well. As discussed in Chapter 8, plants and microorganisms are thought to accelerate weathering by increasing the CO_2 partial pressure in soils and by releasing organic acids that help dissolve rocks. **Vascular plants** have a well-developed stem or trunk for transporting water and nutrients from the ground up to their leaves, and typically a well-developed root system that takes up water and nutrients and helps support the plant. So, one might expect that the spread of vascular plants would have lowered atmospheric CO_2 (Figure 12-12). Unfortunately for this idea, the timing is not right. Vascular plants originated in the Late Silurian and spread widely during the Devonian period, well before the climate

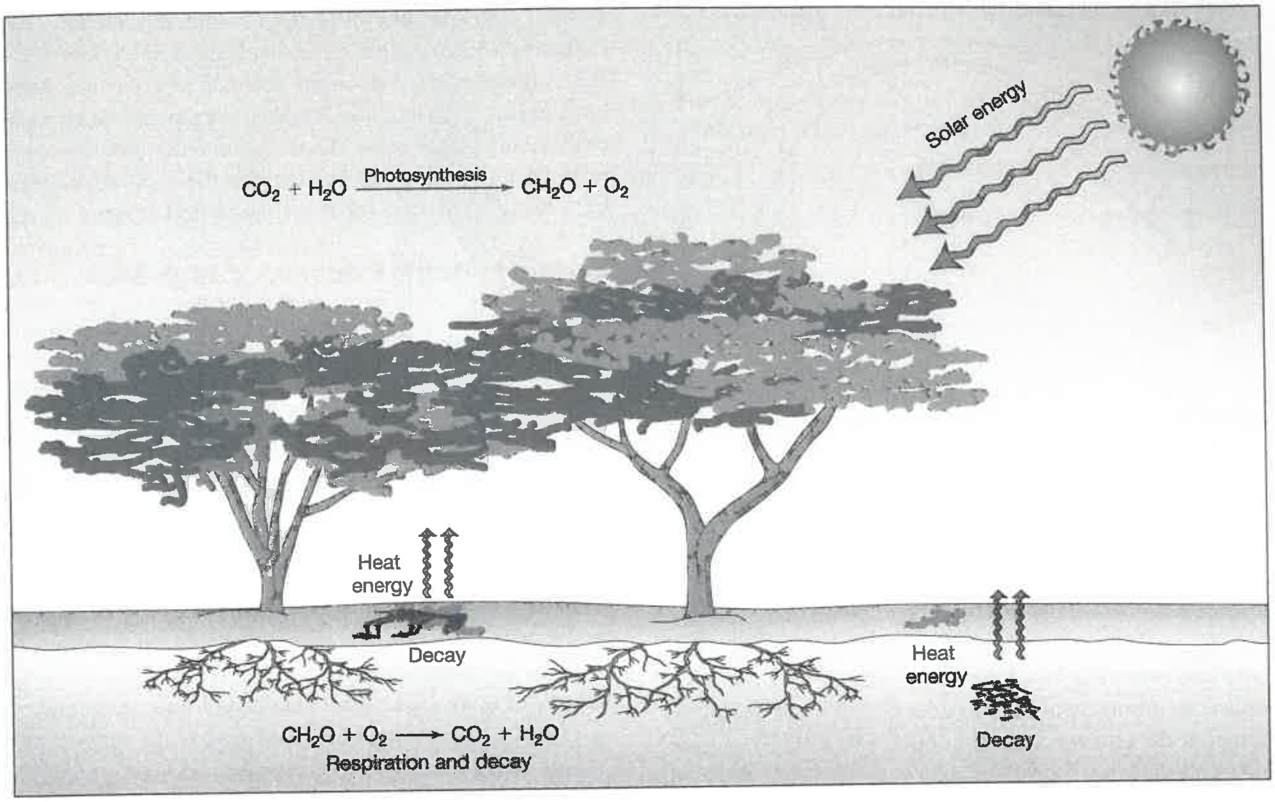


FIGURE 12-12 Land plants enhance the partial pressure of CO_2 in soils through root respiration and decay. (*Source:* From T. McKnight, *Physical Geography: A Landscape Appreciation*, 6/e, 1999. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

began to cool. So, let us consider other factors that might have affected atmospheric CO_2.

A strong clue as to what might have happened is suggested by the carbon isotope record. As discussed in the previous chapter (see Figure 11-16), carbon isotopes indicate that the organic carbon burial rate nearly doubled during the Carboniferous as a result of the formation of large coal beds. In Chapter 11 we suggested that this may have led to an increase in atmospheric O_2. However, it could have led to a decrease in atmospheric CO_2 as well. Robert Berner from Yale University has included this information in a model that tries to predict CO_2 levels during the Phanerozoic (Figure 12-13). As expected, the increased burial of organic carbon during the Carboniferous leads to a substantial drop in atmospheric CO_2 levels. As partial confirmation of Berner's results, the CO_2 concentrations predicted by his model are in approximate agreement with CO_2 levels estimated from paleosol data. This gives us some confidence that the basic idea is correct: during the latter half of its history, Earth's climate has been tightly coupled to atmospheric CO_2 levels.

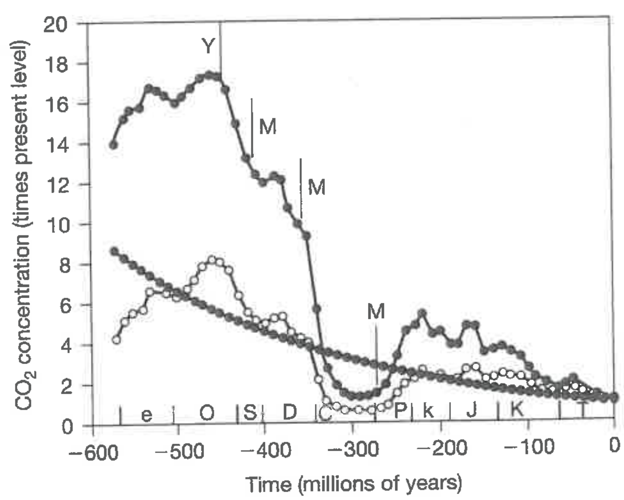


FIGURE 12-13 Phanerozoic atmospheric CO_2 levels predicted by the geochemical cycle model of R. A. Berner. Solid bars represent constraints on CO_2 from paleosol data. (*Source:* R. A. Berner, *Science* 261, 1993, pp. 68–70.)

The Warm Mesozoic Era

The Mesozoic and Cenozoic eras span the past 251 m.y. of Earth history. The Mesozoic, the age of the dinosaurs, is thought to have been considerably warmer than today (Figure 12-11). Because large animals existed during this time, the fossil evidence relating to Mesozoic climate is quite extensive. For example, during the Mid-Cretaceous period, around 100 m.y. ago, lush ferns and alligators resided in what is now Siberia. Dinosaur skeletons have

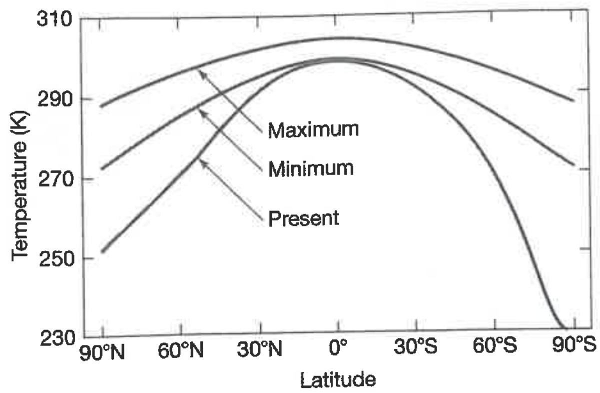


FIGURE 12-14 Estimated limits on longitudinally averaged surface temperatures during the Mid-Cretaceous period, 100 m.y. ago, as compared with today. (*Source:* E. J. Barron and W. M. Washington, "The Carbon Cycle and Atmospheric CO_2," *Geophysical Monograph* 32, AGU, Washington, DC, 1985.)

been recovered from north of the Arctic Circle in Alaska. These and other pieces of evidence indicate that the Mid-Cretaceous climate was on the order of 2 to 6°C (3.6–11°F) warmer at the equator and 20 to 60°C (36–110°F) warmer at the poles (Figure 12-14).

A second type of evidence that indicates that the Mesozoic climate was warm comes from measurement of oxygen isotope ratios in carbonate sediments recovered from deep-sea cores. Although we will postpone detailed discussion of this topic until Chapter 14, oxygen isotopes provide information both on the temperature of the water in which they formed and on the amount of water stored in the polar ice caps. The measured isotope ratios tell us that Mesozoic ocean water was much warmer than today. This is especially true of the deep ocean, which currently has an average temperature of only 2°C. Deep-ocean temperatures during the Mesozoic era were as high as 15°C. Furthermore, the polar ice caps, which today hold enough water to raise sea level by nearly 80 m, appear to have been absent throughout the entire Mesozoic and during the early Cenozoic as well.

What factor or combination of factors was responsible for the extreme warmth of the Mesozoic? In keeping with the discussion of the previous section, suspicion currently centers on higher atmospheric CO_2 levels. Climate models suggest that an increase in CO_2 by a factor of 4 from the present value could explain the warm climate of the Mid-Cretaceous period. This hypothesis is bolstered by paleomagnetic evidence that indicates that the seafloor was spreading faster at that time than it has in the more recent geologic past. Recall that the spreading rate can be estimated by looking at magnetic patterns in the seafloor. Faster spreading rates would have led to faster rates of subduction of carbonate sediments and this, in turn, should have led to increased rates of CO_2 production from carbonate metamorphism. More CO_2 may also have been released by outgassing at the mid-ocean ridges themselves. And higher sea level at that time, itself caused by faster seafloor spreading as well as the absence of polar ice, would have meant that there was less land area available on which to weather silicate rocks. These factors have been included in Berner's models and account for the high atmospheric CO_2 levels predicted for this time period (Figure 12-13).

Carbon Isotopic Evidence of High Mesozoic CO_2 Levels

Our discussion of paleo-CO_2 levels and climate has thus far been based mostly on theory. It has recently become possible, however, to test the theory with data for at least a few time periods in Earth history. Several different methods have been employed. One example, discussed earlier in the chapter, is based on the mineralogy of paleosols. That particular method is not applicable to more recent times because siderite ($FeCO_3$) is not stable under an oxygenated atmosphere. However, a similar type of analysis can be performed on some paleosols using trace carbonates found in the mineral goethite, $Fe(OH)_3$. The paleosol data shown in Figure 12-13 were obtained by this technique.

Another technique for inferring past CO_2 concentrations is based on carbon isotopes. As discussed in the previous chapter, photosynthetic organisms tend to take up ^{12}C faster than they take up ^{13}C. But they do this to a greater extent if CO_2 is relatively abundant in the organism's environment. So, photosynthetically produced organic matter created under high-CO_2 conditions tends to have a low $^{13}C/^{12}C$ ratio, or negative $\delta^{13}C$. If CO_2 is scarce, organisms use whatever isotope is available and the relative isotopic abundances do not change as dramatically.

Carbon isotope abundances have now been measured in sediments of various ages by a number of different research groups. Figure 12-15 shows the difference in $\delta^{13}C$ values between carbonates and organic carbon. The story that has emerged is consistent with that told above: organic matter from sediments of Mesozoic age contains less ^{13}C than organic matter from sediments deposited during the last 20 m.y., indicating that atmospheric CO_2 levels were probably higher at that time. Although there is still considerable disagreement concerning how much CO_2 has decreased, the idea that CO_2 is a main driver of climate on these time scales has received additional support.

Other Possible Influences on Mesozoic Climate

Although higher atmospheric CO_2 levels can account for the overall warmth of the Mesozoic climate, this mechanism cannot by itself explain the extremely small latitudinal temperature gradient at that time. The equator-to-pole temperature contrast during the Mid-Cretaceous period was only 20–30°C, as compared to 50–60° today. Part of this difference can be explained by the absence of polar ice at that time. Recall that ice cover interacts with climate

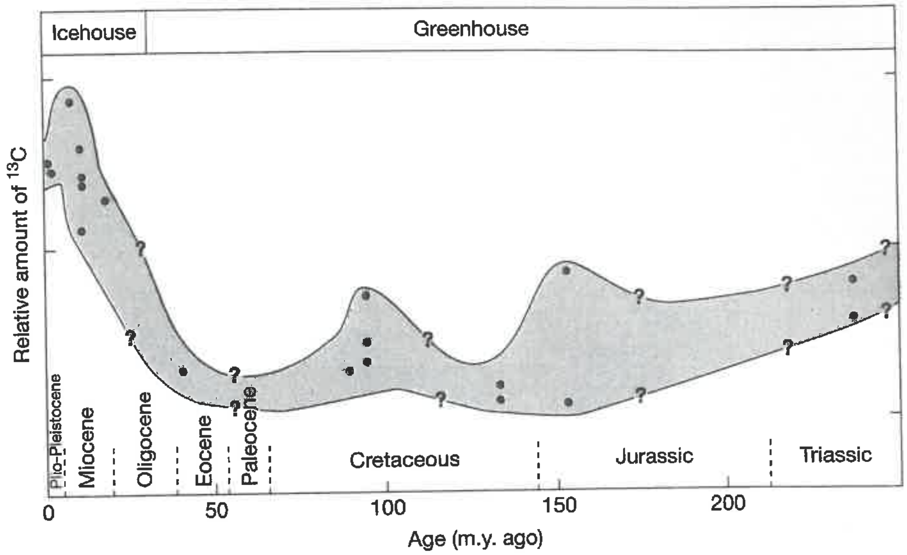


FIGURE 12-15 Carbon isotope data for the past 150 m.y. The vertical scale is a measure of the difference in ^{13}C content between carbonates and organic carbon. Large differences in ^{13}C correspond to lower atmospheric CO_2 values. The data are consistent with a decrease in CO_2 since the Mid-Cretaceous period. (*Source:* B. Popp et al., *American Journal of Science* 289, 1989, pp. 436.)

by way of a strong, positive feedback loop. Removal of the ice caps would cause a large decrease in the albedo of the polar regions that, in turn, should cause them to warm significantly.

Calculations with climate models suggest, however, that ice–albedo feedback alone is not sufficient to explain the extreme warmth of the Mid-Cretaceous poles. It seems likely that the atmosphere–ocean system was for some reason more effective in transporting heat from the equator to the poles than it is today. One possibility is that the thermohaline circulation of the oceans ran backward at that time: warm, but highly saline, deep water formed at low latitudes welled up near the poles, where it then warmed the climate through evaporation. Unfortunately, no one has yet demonstrated that this mechanism could work. A second possibility is that the tropical Hadley circulation extended further poleward than it does today. Because Hadley cells are very efficient at transporting heat, this mechanism could explain the low latitudinal temperature gradient. Moreover, it could explain the apparent absence of subfreezing temperatures in continental interiors (Siberia, for example) during the long polar night. Again, however, no one has demonstrated that this mechanism is dynamically feasible.

One of the authors of this book (LK), together with David Pollard (mentioned earlier), have proposed yet another mechanism for additional warmth during these already warm periods: reduced biological production of the cloud condensation nucleus dimethyl sulfide, leading to fewer and thinner clouds (see Chapter 11). Warm intervals would be less biologically productive because of reduced nutrient supply by upwelling; the strong thermocline (warmer surface waters above cool deep waters) would provide an effective barrier to upwelling by stabilizing the water column density structure. This mechanism leads to overall amplification of global warming and an intensification of warming near the poles. Perhaps the Cretaceous sky not only contained higher greenhouse gas concentrations but was less cloudy as well!

Cooling during the Cenozoic Era

Starting about 80 m.y. ago, Earth's climate began to cool (except for a short-lived warming during the early Eocene period). The initial decrease may simply have been caused by a decrease in mid-ocean ridge spreading rates, leading to a reduction in atmospheric CO_2. However, the cooling trend accelerated around 30 m.y. ago during the Oligocene epoch in a way that does not correlate with the spreading-rate data. Thus, paleoclimatologists have searched for other explanations for the observed cooling. One intriguing theory, suggested by Maureen Raymo, then at the Massachusetts Institute of Technology, and William Ruddiman of the University of Virginia, is that the carbonate–silicate cycle was perturbed by plate tectonics, but by a mechanism that differs from those discussed previously.

During the Mesozoic and early Cenozoic eras, India was a separate continent drifting slowly toward Asia. The two continents collided around 40 m.y. ago, a process that is still continuing today. The collision created a gigantic chain of mountains (the Himalayas) and a huge area of uplifted terrain called the Tibetan Plateau (Figure 12-16). The Himalayan Mountains provided fresh, readily erodable surfaces on which silicate weathering could proceed rapidly. At the same time, the uplift of the Tibetan Plateau created seasonal rainfall (the southeast Asian monsoon), which provided the water needed for weathering to occur on the face of the Himalayan range. The combination of these factors may have accelerated silicate weathering rates over a substantial portion of Earth's surface, thereby helping bring atmospheric CO_2 concentrations down to the relatively low levels that prevail today.

The point to be drawn from this discussion is that plate tectonics probably does influence climate over long

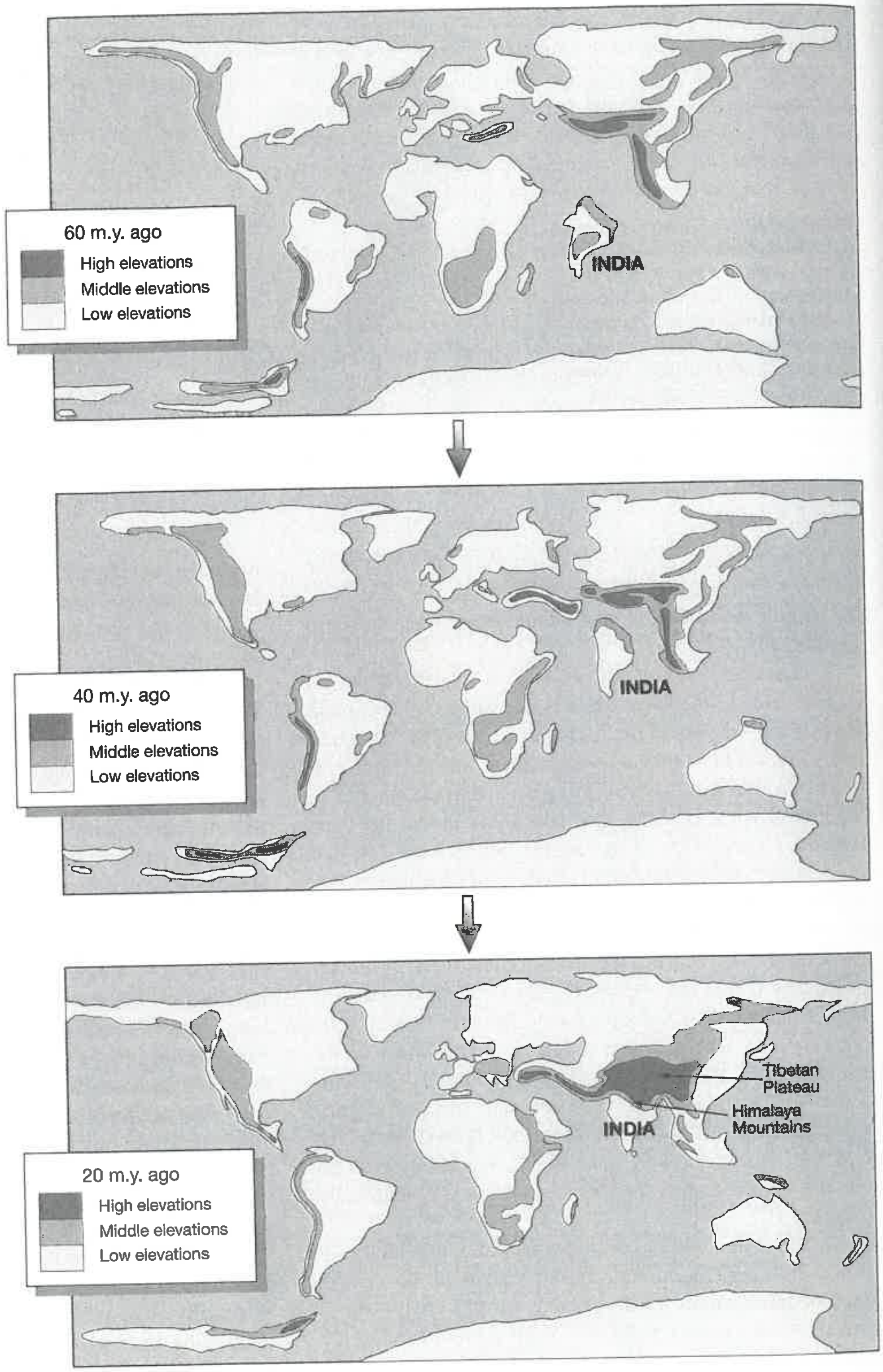


FIGURE 12-16 The collision of India with Asia, shown in a series of paleogeographic maps for the Paleocene (60 m.y. ago), the Eocene (40 m.y. ago), and the Miocene (20 m.y. ago) eras. (*Source:* E. J. Barron, *Paleogeography, Paleoclimatology, Paleoecology* 50, 1985, p. 45.)

time scales, but not necessarily in the way that geologists have traditionally imagined. Continental-scale glaciers can indeed grow when landmasses drift close to the poles but only if climatic conditions are ripe for such a development—specifically, if atmospheric CO_2 concentrations are relatively low. The main influence of plate tectonics on climate appears to be indirect: by changing the way in which the carbonate–silicate cycle operates, plate tectonics helps modulate atmospheric CO_2 levels. This, in turn, affects climate by way of the greenhouse effect. Such changes, in combination with the long-term increase in solar luminosity, can account for the main features of the long-term climate record.

Chapter Summary

1. During the early parts of Earth's history, the faintness of the young Sun must have been offset by higher concentrations of greenhouse gases in the atmosphere.
   a. CO_2 may have dominated at first, but CH_4 was probably an important greenhouse gas as well, once it was being produced by methanogenic bacteria.
   b. The combination of CO_2 and CH_4 kept Earth's climate relatively warm until the early Paleoproterozoic era.
   c. The rise of atmospheric O_2 at ~2.4 b.y. ago eliminated most of the CH_4 and thereby triggered the Paleoproterozoic glaciation.
   d. That said, CH_4 could still have remained an important greenhouse gas during much of the Proterozoic eon, provided that the deep oceans remained anoxic. A decrease in CH_4 toward the end of this eon, perhaps triggered by a corresponding rise in atmospheric O_2, could have provided the trigger for the Neoproterozoic Snowball Earth glaciations.
2. During Earth's more recent history, climate has been largely determined by the balance between increasing solar luminosity and decreasing atmospheric CO_2.
   a. During most of this time, climate has been kept within moderate bounds by the negative feedback associated with the carbonate–silicate cycle.
   b. This stabilization mechanism has broken down temporarily, however, at least three times, resulting in global glaciations near the beginning and the end of the Proterozoic. Life survived these Snowball Earth episodes by mechanisms that are currently being debated.
3. During the last 500 m.y. of Earth's history, climate has alternated between periods of warmth (the Mid-Cretaceous) and periods of cold (the Late Ordovician, Permo-Carboniferous, and Pleistocene glaciations). Earth is currently in a moderately cold state—a brief interglacial period within the Pleistocene glacial epoch. Variations in atmospheric CO_2 levels caused by changes in plate tectonics and by biological innovations can explain the broad features of Phanerozoic climate history.

Key Terms

anti-greenhouse effect
cap carbonates
corona
dropstones
ethane
geothermal heat
glacial striations
ice-rafting
methanogenesis
Mie scattering
moraines
neutrinos
paleoclimate
polymerize
Rayleigh scattering
reduced gas
scattering
solar wind
sulfate-reducing bacteria
thermophiles
tillite
vascular plants

Review Questions

1. Why does the Sun get brighter with time?
2. How might the carbonate–silicate cycle have helped solve the faint young Sun problem?
3. Why is methane thought to have been an important greenhouse gas during the Archean era?
4. What triggered the Huronian glaciation at 2.4 b.y. ago?
5. What types of geologic evidence are used to infer past glaciations?
6. How many separate episodes of glaciation have occurred during Earth's history?
7. What types of geologic evidence support the Snowball Earth model for the Late Precambrian glaciations?
8. How are carbon isotopes used to infer past atmospheric CO_2 concentrations?
9. How are atmospheric CO_2 levels affected by the presence of land plants?
10. What mechanisms might explain the warm climate of the Mesozoic era? How might the equator-to-pole temperature gradient have been reduced?
11. Why did climate cool during the past 40 million years?

Critical-Thinking Problems

1. Evidence for low-latitude glaciation is found at both 0.6 Ga and 2.4 Ga. These are two of the three possible Snowball Earth events mentioned in the text. (We will neglect the event at 0.75 Ga because it is similar to the first one.) Your job is to estimate how thick the ice was at those times.
   a. The variation in solar luminosity with time can be approximated by the following formula (derived by fitting the results of a computer model of the Sun's evolution):

$$S = \frac{S_0}{1 + 0.4(t/4.6)}$$

where S = solar flux at time t

S_o = 1370 W/m^2 = present solar flux
t = time in Ga (billions of years before present).

Calculate the solar flux at 0.6 and 2.4 Ga both in W/m^2 and as a percentage of its current value.

b. As we have learned previously, the effective radiating temperature of Earth can be found from the formula

$$\sigma T_e^4 = \frac{S}{4}(1 - A),$$

where A is the planetary albedo and $\sigma(= 5.67 \times 10^{-8}$ $W/m^2/K^4$) is the Stefan–Boltzmann constant. Calculate T_e at 0.6 and 2.3 Ga, assuming an albedo of 0.65 (the value for clean ice and snow). Then, calculate the global average surface temperature, T_s, assuming that the atmospheric greenhouse effect, ΔT_g, was the same as today (33 K). Recall that $T_s = T_e + \Delta T_g$.

c. The conductive heat flow through ice is given by

$$F = \frac{\lambda \Delta T}{\Delta z},$$

where λ (= 2 W/m/K) is the thermal conductivity of ice, ΔT is the temperature difference between the top and bottom of the ice layer, and Δz is the thickness of the layer. We know that the current geothermal heat flux, F, is about 0.09 W/m^2. Assume that F had this same value at 0.6 Ga, but was twice as high at 2.4 Ga. Assume also that the top of the ice is at temperature T_s and that the water below the ice has a temperature of −2°C (the freezing point of seawater). What is the thickness of the ice at both 0.6 and 2.4 Ga?

2. Suppose now that the ice is transparent enough so that some sunlight makes it through. Let's see how that would change the ice thickness.

a. Calculate the globally averaged solar flux incident on Earth's surface during the Neoproterozoic glaciation (0.6 Ga).

b. The solar flux at the equator is about 20% higher than the global average value. Suppose that 10% of this incident sunlight makes it through the ice. How thin must the ice layer be in order to conduct this heat back out so that the water temperature remains constant? (Use the formula from Problem 1c.)

c. Is the ice thickness calculated in part (b) consistent with the assumption that it would transmit 10% of the sunlight through it? Determine this by consulting Box Figure 12-2.

Further Reading

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