Annu. Rev. Astron. Astrophys. 1994. 32: 83–114

THE GOLDILOCKS PROBLEM: Climatic Evolution and Long-Term Habitability of Terrestrial Planets

Michael R. Rampino

Department of Earth System Science, New York University, New York, New York 10003

Ken Caldeira

Global Climate Research Division, Lawrence Livermore National Laboratory, Livermore, California 94551

KEY WORDS: atmospheric evolution, greenhouse effect, Mars, Venus

INTRODUCTION

Why is Venus too hot, Mars too cold, and Earth "just right" for life? (The allusion to the fairy tale involves the three bowls of porridge belonging to Papa Bear, Mama Bear, and Baby Bear—one too hot, one too cold, and one just right—tested by a hungry Goldilocks.) A simplistic answer might be that a planet's surface temperature is to a large extent a function of its distance from the Sun, and Earth just happens to be at the "right" distance for comfortable temperatures and liquid water. However, this is far from the whole story.

The Goldilocks Problem involves the early history of the planets and the evolution of their atmospheres. Its solution must also take into consideration the long-term evolution of the Sun, and hence the so-called faint young Sun problem, that is, the fact that the early Earth was apparently warm enough for liquid water despite the 25–30% lower luminosity of the early Sun (Newman & Rood 1977; Gough 1981). Had Earth been too cold initially for liquid water to exist on its surface, the resulting icy planet would have had a high albedo or reflectivity, lowering temperatures further, and might have become irreversibly ice-covered—the "white Earth catastrophe" (Caldeira & Kasting 1992a). Yet

0066–4146/94/0915–0083$05.00

evidence exists that liquid water has been abundant on Earth for at least the last 3.8 billion years.

The white Earth catastrophe might be averted through geologic activity that provides continued outgassing of CO_2, thereby warming the planet, and eventually melting the ice. But could too much CO_2 produce surface conditions too hot for liquid water, arresting the rock weathering reactions that act to remove CO_2 from the atmosphere, and creating a dense, hot CO_2-rich atmosphere, such as present on Venus today?

Many scientists have stressed the importance of the origin and evolution of life on Earth in biogeochemical cycling of carbon and in causing important changes in atmospheric composition over the last 4 billion years. Proponents of the Gaia hypothesis (Lovelock & Margulis 1974; Lovelock 1979, 1989) go further in claiming that life itself has managed to maintain surface conditions on Earth within a fairly narrow window through a series of negative feedbacks involving greenhouse gases, cloud albedo, and other factors.

Thus, despite its trivial sounding name, the Goldilocks Problem touches on major questions in stellar evolution, the origin of planets, geologic activity, the origin and evolution of life, biogeochemical cycles, climate modeling, the search for extraterrestrial intelligence, and on the place of life and human existence in the Cosmos. Recent reviews have emphasized various aspects of the problem (e.g. Prinn & Fegley 1987, Kasting 1989, Priem 1990, Pollack 1991, McKay 1991, Pepin 1991, Hunten 1993); this review seeks to clarify the Goldilocks Problem within the larger questions of stellar and planetary evolution and planetary habitability, and to discuss some of the most recent work on the problem.

PLANETARY SURFACE TEMPERATURES

The effective temperature (T_e) of a planet's surface, determined by the planet's distance from the Sun and its surface albedo, is defined as:

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

where S is the amount of solar insolation at the planet's distance from the Sun, A is the planet's albedo, and σ is the Stefan-Boltzmann constant.

Earth, Mars, and Venus have atmospheres that contain the important greenhouse gases, H_2O vapor and CO_2, which absorb outgoing long-wave radiation, and thus warm the planet's surface (Table 1).

Therefore, the surface temperatures (T_s) of the these planets may be defined as the sum of their effective temperatures (T_e) plus the greenhouse effect (δT) provided by the greenhouse gases in the atmosphere ($T_s = T_e + \delta T$) (Goody & Walker 1972, Pollack 1991). Table 2 shows the values of T_e and T_s for Venus, Earth, and Mars.

Table 1 Main constituents of terrestrial planetary atmospheres

Planet	Atmospheric pressure (bars)	Major constituents[a]
Venus	90	$\mathbf{CO_2}$ (0.96), N_2 (0.035), $\mathbf{H_2O}$ (4×10^{-5})
Earth	1	N_2 (0.77), O_2 (0.21), $\mathbf{H_2O}$ (~0.01), Ar (0.009), $\mathbf{CO_2}$ (3.5×10^{-4})
Mars	0.006	$\mathbf{CO_2}$ (0.95), N_2 (0.027), Ar (0.016), $\mathbf{H_2O}$ (3×10^{-4})

[a] Numbers in parentheses show the fractional abundance, by number, of the major atmospheric constituents. Boldface indicates greenhouse gases. Other gases can contribute to greenhouse warming through pressure broadening of the absorption bands of greenhouse gases.

From Table 2, it can be seen that our original conjecture that the Goldilocks Problem could be solved merely by taking into account the differences in the planetary distances from the Sun does not work—T_e for *all three* planets is below the freezing point of water. On Earth, the comfortable conditions for life are made possible by the 33 K of greenhouse warming supplied by H_2O and CO_2 gases in the planet's atmosphere. Note also that although Venus is the closest to the Sun of the three planets, its higher surface temperature is not a direct result of increased solar insolation—cloud-covered Venus absorbs only slightly more solar radiation than barren Mars as a result of the great differences in their present albedoes. The greenhouse effect of Venus' dense (90 bar) CO_2 atmosphere adds 521°C of warming. The effectiveness of greenhouse heating on Venus is increased by the fact that greenhouse gases can absorb thermal radiation in their weaker transitions as total atmospheric mass increases, thus the fraction of thermal IR absorbed increases with a planet's surface pressure (Pollack 1991).

Table 2 Variation of effective and surface temperature on three planets showing the influence of greenhouse gases[a]

Planet	Atmosphere pressure (atm)	Greenhouse gases	Orbit (AU)	Solar constant (W m^{-2})	Albedo (%)	Effective temp (K)	Surface temp (K)	Greenhouse warming (°C)
Venus	90	CO_2	0.723	2620	76	229	750	521
Earth	1	H_2O, CO_2	1.00	1368	30	255	288	33
Mars	0.006	CO_2	1.524	589	25	210	218	8

[a] Note that the relationship among the solar insolation values (S/S_0) taking Earth = 1.00, are Venus = 1.91, and Mars = 0.43.

It is important at this point to note that the amount of CO_2 at the surface of Earth is also very high, equivalent to ~60 bar, but on Earth the CO_2 is tied up in carbonate rocks as solid $CaCO_3$ (Ronov & Yaroshevsky 1976, Holland 1978), and additional CO_2 is present in the upper mantle. This suggests that by whatever method CO_2 accumulates in a planet's atmosphere (e.g. outgassing from the interior, delivery by impactors during the planetary accretion process), Venus and Earth may have received about the same complement of the gas. The difference is that on Earth the CO_2 became locked up in surface rocks, whereas on Venus the CO_2 has remained in the atmosphere, creating the extreme greenhouse conditions on the planet today. The question is: Why did things turn out so badly for our sister planet?

TOO HOT: THE RUNAWAY GREENHOUSE

Venus today is an uninhabitable inferno with a dense carbon dioxide atmosphere and a surface temperature of 750 K (477°C)—hot enough to melt lead. The atmosphere is also exceedingly dry—Donahue & Hodges (1992) recently provided evidence that the H_2O mixing ratio is only $\sim 3 \times 10^{-5}$ in the Venus atmosphere. The high temperatures and lack of water on Venus are commonly attributed to a "runaway greenhouse," in which H_2O was lost, and CO_2 built up in the atmosphere to the high levels seen today.

The idea of a runaway greenhouse was suggested by Hoyle (1955), and has been further developed by a number of workers including Sagan (1960), Gold (1964), Dayhoff et al (1967), Ingersoll (1969), Rasool & DeBergh (1970), Pollack (1971), Goody & Walker (1972), Walker (1975), Watson et al (1981), Matsui & Abe (1986a,b), Abe & Matsui (1988), Kasting (1988, 1989, 1991a), Durham & Chamberlain (1989), and Tajika & Matsui (1992). Early ideas are well summarized in Walker (1977), Henderson-Sellers (1978), and Henderson-Sellers & Cogley (1982). These early considerations of the runaway greenhouse are exemplified in Figure 1 from Rasool & DeBergh (1970) (see Goody & Walker 1972).

The three curves in the figure represent the evolution of the surface temperatures of the terrestrial planets (starting with no atmosphere and an albedo like that of present-day Mars) as water vapor, either with or without CO_2, is released into the atmosphere by planetary volcanic outgassing. The curves show how surface temperatures increase as a result of the greenhouse effect, as water vapor accumulates in the atmospheres of Venus, Earth, and Mars. On Mars and on Earth, the increase is halted when the water vapor pressure is equal to the saturation vapor pressure, and freezing or condensation occurs. The Martian temperatures are so low that gases do not accumulate in the atmosphere for very long before the pressure reaches the saturation pressure of ice. This

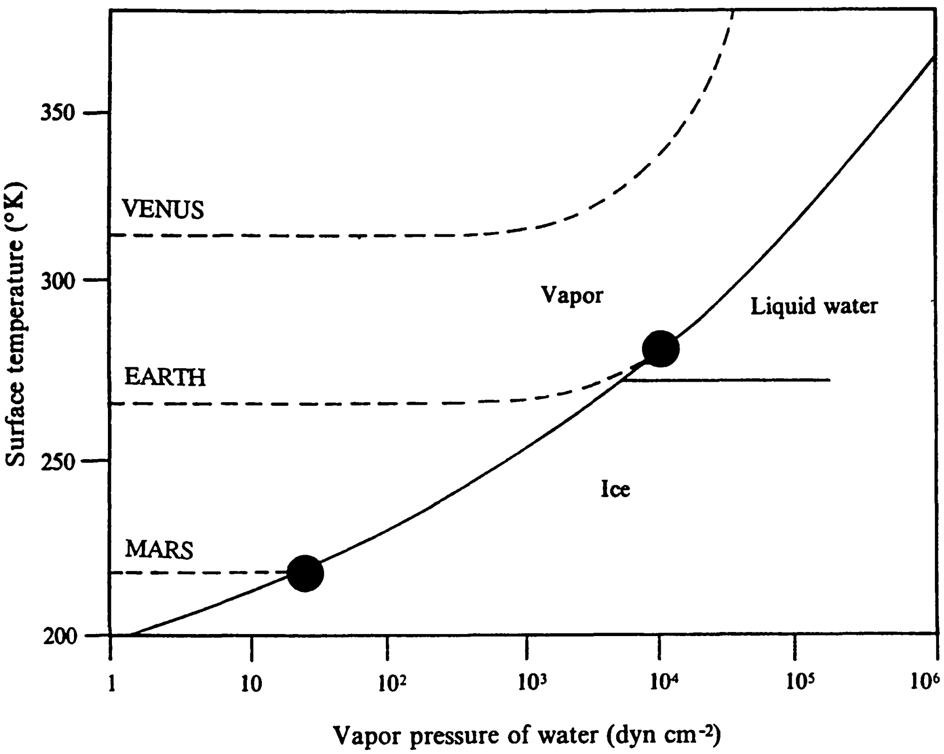


Figure 1 An early interpretation of the runaway greenhouse effect. The dashed curves show how surface temperatures increase, due to the greenhouse effect, as water vapor accumulates in the atmospheres of Venus, Earth, and Mars. On Mars and Earth, the increase is halted when the water vapor pressure is equal to the saturation vapor pressure, and either freezing or condensation occurs. Temperatures on Venus are higher because Venus is closer to the Sun, and saturation is never achieved. Therefore, temperature runs away. Note that the temperatures on the left-hand axis are not the same for Earth and Venus as the effective temperatures in Table 1, as different planetary albedoes were used. (After Goody & Walker 1972.)

produces a thin atmosphere with a small greenhouse effect associated with low atmospheric H_2O vapor pressure.

Because Venus is closer to the Sun, initial surface temperatures were higher. According to this scenario, outgassing of H_2O and CO_2 produced a greenhouse effect that increased the surface temperature, which, in turn, increased the saturation vapor pressure of water in the atmosphere. The greenhouse effect of the increased water vapor is so great that saturation was never achieved, and condensation never occurred. As more water vapor is released, a strong positive feedback develops, temperatures increase further, the saturation vapor pressure of water increases further, etc, and the temperature eventually runs away. In this picture, oceans never formed on Venus; the water has always existed as water vapor in the atmosphere.

The presence of H_2O in the upper atmosphere inevitably leads to loss of water by photodissociation (Goody & Walker 1972). The disappearance of liquid water prevents the so-called Urey silicate-rock weathering reactions, e.g.:

$$CaSiO_3 + CO_2 \Rightarrow CaCO_3 + SiO_2$$

which act to remove CO_2 from planetary atmospheres, from taking place (Walker et al 1981). These reactions occur in the fluid phase, or at the fluid/solid interface, and so require liquid water. Therefore, the continued outgassing of CO_2 will further increase the planet's surface temperature in a positive feedback loop, leading eventually to an extremely high CO_2 atmosphere, such as seen on Venus today (Kasting 1989). In size, Venus is Earth's close twin (equatorial diameter of 12,104 km vs 12,756 km for Earth), and the planet's bulk should have efficiently sealed in the heat produced by accretionary impacts and radioactive decay. Venus is apparently volcanically active, although the presence of some process resembling terrestrial plate tectonics remains debatable (Head et al 1992).

It is now apparent that the early analyses of the runaway greenhouse had two problems:

1. They neglected convection in the lower atmosphere, which would change the vertical profile of the mixing ratio of water vapor in the planet's atmosphere, and reduce the lapse rate (the rate of temperature decrease with altitude). This greatly reduces the magnitude of the greenhouse effect. The elevated temperatures at high altitudes mean that H_2O mixing ratios are relatively high in the upper atmosphere; and

2. Although the low abundance of non-radiogenic rare gases in the atmospheres of the three planets indicates that they did not acquire a primordial atmosphere directly from the solar nebula, the current models of rapid planet formation suggest that planets start out with dense atmospheres from degassing of impactors during accretion, and/or the addition of volatile-rich cometary material in the latter stages of accretion (e.g. Matsui & Abe 1986a,b; Abe & Matsui 1988; Chyba 1990).

Another way of looking at the runaway greenhouse is to consider the critical value of the solar flux incident at the top of a planet's atmosphere above which liquid water cannot exist at the surface (Komabayashi 1967, Ingersoll 1969, Hart 1978, Kasting 1991a, Nakajima et al 1992). For example, if Earth was by some means pushed closer and closer to the Sun (or the Sun's luminosity increased), we would anticipate that at some point, the oceans would be vaporized, and the planet would be enveloped in a dense steam atmosphere. The amount of water in Earth's oceans is 1.4×10^{24}g, which means that the surface pressure of this atmosphere would be ~270 bar, ~50 bar greater than the pressure at the critical point of water.

Nearly all solar models indicate that the Sun has been getting more luminous with time, by ~30% since the Solar System formed (but see Gilliland 1989), because as the Sun converts hydrogen to helium the Sun's core becomes denser and hotter, increasing the rate of thermonuclear fusion (Newman & Rood 1977). The luminosity (S) of the Sun on the main sequence as measured at Earth using a standard solar model (Sackman et al 1990) can be approximated by

$$S(t) = \left(1 - \frac{0.38t}{\tau_0}\right)^{-1} S_0$$

for the interval -4.5 billion years $< t <$ 4.77 billion years (Caldeira & Kasting 1992b). Here t is time expressed as years from the present, $\tau_0 = 4.55$ billion years, and the subscript 0 refers to present day values ($S_0 = 1368\ \mathrm{Wm}^{-2}$).

Thus, for the Goldilocks Problem of planetary habitability, especially with changing solar output over time, the most important aspect of a runaway greenhouse is the critical solar flux required to trigger it. How much additional solar insolation would be required to turn Earth into Venus? Or alternately, how much closer to the Sun would Earth need to have been to put it on the path to a runaway greenhouse? How lucky were we?

This critical question has been discussed in a number of papers (e.g. Hart 1978, Rasool & DeBergh 1970, Kasting 1988). Rasool & DeBergh (1970) calculated that the water in the oceans would never have condensed if Earth were formed 4 to 7% (6 to 10×10^6 km) closer to the Sun, and Hart (1978) calculated about 5% closer (0.95 AU). More recently, using a radiative-convective climate model and treating solar flux as a variable, Kasting (1988) made detailed estimates of this energy threshold for a fully saturated, cloud-free atmosphere, where F_{IR} = outgoing IR, and F_{s} = incoming solar radiation:

$$F_{\mathrm{IR}} = \frac{S}{S_0}(F_{\mathrm{s}}).$$

Assuming a surface temperature and vertical temperature profile, and calculating F_{IR} and F_{s}, the critical solar flux in his model was 1.4 S_0. Abe & Matsui (1988) performed a similar calculation for a case in which part of the energy required to trigger a runaway greenhouse was derived from infalling planetesimals during planetary accretion, and the results agreed well with those of Kasting (1988). These findings are in agreement with the inferred history of Venus, where, because of the Sun's lower early luminosity, the solar flux (now 1.91 S_0) would have been ~1.3 to 1.4 S_0 early in the history of the Solar System.

AN ALTERNATIVE: THE MOIST GREENHOUSE

Kasting and co-workers (1984, 1988) proposed an alternative model to the runaway greenhouse, the "moist greenhouse," in which Venus could have lost its water while maintaining liquid oceans. The concept comes from studies suggesting that the vertical distribution of water vapor in a planet's atmosphere should be strongly correlated with its mixing ratio near the surface (Ingersoll

1969). When water vapor is a minor constituent of the lower atmosphere, the concentration declines rapidly with altitude throughout the convective region as a consequence of condensation and rainout (Kasting 1988). This takes place in the present terrestrial atmosphere, where the mixing ratio for water vapor drops from ~0.01 near the surface to $\sim 3 \times 10^{-6}$ in the lower stratosphere. The upper troposphere provides a so-called cold trap which prevents much water vapor from entering the stratosphere. Ingersoll (1969) showed that in a convecting atmosphere with a water vapor mixing ratio > 0.1, the release of latent heat by the condensing water keeps the lapse rate at a small value so that the atmospheric temperature decreases very slowly with height. The almost constant mixing ratio means that water can remain at significant levels even at high altitudes, where it can be effectively dissociated by solar ultraviolet (UV) with subsequent loss of hydrogen to space. As long as sufficient solar extreme ultraviolet (EUV, with $l \leq 100$ nm) energy is available to cause the escape, H should escape at about the diffusion-limited rate (Hunten 1973).

The work of Kasting et al (1984) and Kasting (1988) suggested that hydrogen escape becomes very rapid for incident solar fluxes exceeding 1.1 S_0. The calculations indicate that Venus could have lost most of its water without ever experiencing true runaway greenhouse conditions. In a plot of $S/S_0(S_{\text{eff}})$ against surface temperature, it seems that early Venus may have been just at the transition region between runaway and moist greenhouse situations (Figure 2). The solar flux at the orbit of Venus early in the history of the Solar

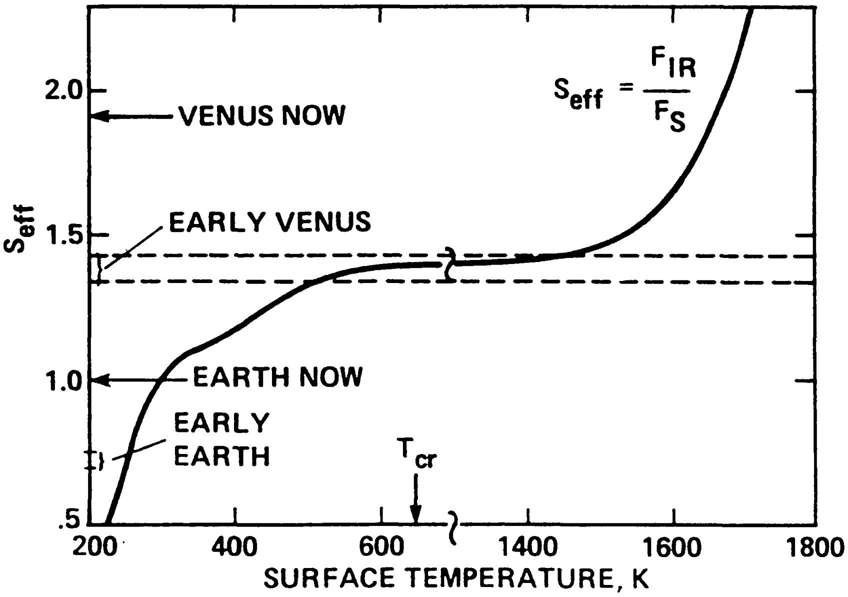


Figure 2 Effective solar constant S_{eff} vs surface temperature. The horizontal dashed lines represent estimates of the solar flux at the orbit of Venus ~4.6 billion years ago. Note that there is a break in the horizontal scale between 700 and 1300 K. (After Kasting 1988.)

System (~1.3 to 1.4 S_0) is so close to the minimum calculated for a runaway greenhouse (1.4 S_0), that Kasting (1988) suggested that the effect of increased clouds, which would reflect more solar radiation back to space, would have tipped the balance in favor of the moist greenhouse. These modeling efforts suggested that Earth could experience a moist greenhouse catastrophe when solar insolation has increased by ~10% (although this may be prevented by negative cloud feedback)

WATER LOSS FROM VENUS

The present Venus atmosphere has only 10^{-5} the water in Earth's oceans. Either Venus was originally water poor, or the planet has lost much of its water. Lewis (1982) suggested that, according to the equilibrium condensation model of planet formation, Venus would have originally been rather dry. This is because the material that condensed out of the solar nebula at Venus' distance from the Sun is predicted to have had much less of the hydrous silicate phases than the condensate at greater distances from the Sun. However, two problems exist with this model: 1. It is apparently kinetically impossible to form these hydrous phases at the densities of the solar nebula in the relatively short time (10^8 yr) in which the planets formed (Prinn & Fegley 1987); the water should have condensed out of the solar nebula primarily as water ice; and 2. Wetherill (1986) has made a good case that the planets did not form where they are now. His calculations suggest that planetary eccentricities were pumped up to high values during the latter stages of accretion, and that an exchange of material from different parts of the solar nebula took place. The cumulative crater density on the inner planets, and the modern ideas on the formation of the Moon by the collision of a large (Mars-sized) body with Earth support this picture (Wetherill 1986), and if this is correct, then Venus could have initially received a significant amount of water, possibly as much water as Earth.

One measure of the amount of water lost from the atmosphere of Venus is the D/H ratio of the present water vapor, which is ~2.4×10^{-2} (Donahue et al 1982, McElroy 1982, DeBergh et al 1991, Donahue & Hartle 1992). This is ~150 times Earth's D/H ratio of 1.6×10^{-4}. Donahue et al (1982) originally interpreted this difference as indicating that Venus once had ~100 times as much water as it does at present. Since the present water abundance on Venus is only 0.0014% that of Earth, the conclusion is that Venus started out with only about 0.1% of Earth's present water budget. However, if some deuterium was lost along with hydrogen during the escape process, then the amount of water that was lost could have been orders of magnitude greater, and Venus could have started out with as much water as Earth (Donahue et al 1982, Kasting & Pollack 1983).

Taking the case of Rayleigh fractionation—escape with no resupply—the time constant τ for the evolution of the D/H ratio in the Venus atmosphere is

$$\tau \sim R/(f\phi),$$

where R is the vertical column abundance of water vapor in the atmosphere, ϕ is the hydrogen escape rate, and f is the D/H fractionation factor (the relative efficiency of D escape compared to H escape):

$$f = \frac{\phi_D/N_D}{\phi_H/N_H}$$

where ϕ is the escape rate, and N is the column abundance in the atmosphere (Grinspoon 1987).

Recently, Gurwell & Yung (1992) estimated a value of f of 0.125, which is ten times greater than the canonical value of 0.013. This higher fractionation factor takes into account reactions between H and hot Oxygen (O^0). Calculations using the new fractionation factor (Grinspoon 1993) give $R_O/R_t = 300$, equivalent to ~0.3% of a terrestrial ocean, i.e. Venus could have originally had ~10 meter layer of planetary water. The escape time for water τ_{esc} from the planet could have been ~190–240 million years (a minimum value, since f should have been higher early in Venus' history when hydrodynamic escape was possible).

The picture of an early wet Venus was earlier challenged by Grinspoon (1987) and Grinspoon & Lewis (1988), who pointed out that the present D/H enrichment on Venus might be explained if the water abundance in the atmosphere were in a steady state, with loss of water by photodissociation and hydrogen escape balanced by continued influx of water-rich comets. However, it has been suggested that the steady-state picture cannot explain the present D/H ratio of the Venus atmosphere (J. F. Kasting, personal communication 1993).

Venus could have outgassed H_2O that was more enriched in deuterium than that outgassed on Earth. However, the homogenization of planetesimals during accretion (Wetherill 1986) and the fact that the D/H ratio of chondritic meteorites and of Halley's Comet (Eberhardt et al 1987) are similar to the terrestrial value, suggests a single Solar System value. Another possibility is that Venus might have had an ocean at one time that was enriched in D/H by hydrodynamic escape. If Venus had some process like plate tectonics, some of this water could have gone into hydrated seafloor and would have been subducted. Venus could now be outgassing this D-rich water, which is further enriched in D by present-day nonthermal escape mechanisms (J. F. Kasting, personal communication 1993). It is also worth noting that the surface temperature of a planet may affect convection patterns and the chemical structure of the mantle, providing some feedback between outgassed CO_2 and H_2O, planetary surface temperatures, and geologic activity that have yet to be fully explored (Ogawa 1993).

Getting rid of the last few bars of water presents a problem in both the runaway greenhouse and moist greenhouse models (Kasting 1988), but if plate tectonics, or some process like it, exists or once existed on Venus, the remaining water may have gone into the hydrated crust, which was then subducted and returned to the planet's mantle. Volcanic outgassing apparently continued on Venus, increasing atmospheric pCO_2 and the planet's greenhouse effect. Outgassed SO_2 accumulated in the atmosphere as clouds of sulfuric acid. The end result is the present hot and arid conditions on Venus.

Durham & Chamberlain (1989) recently asked the reverse question: What levels of CO_2 are required to produce the climatic conditions thought to have existed on the three planets ~4.0 billion years ago? From a comparative analysis of the planets, making a number of assumptions about the early outgassing history and using a one-dimensional radiative/convective model, they estimated that the Venus atmosphere became unstable at $\geq$ 11 bars of CO_2, whereas early Earth and Mars had stable atmospheres that would not have gone into a moist or runaway greenhouse even if the pCO_2 were equal to 100 bars on Earth. The surface pCO_2 on early Mars could have been ~1.3 to 2.1 bars, whereas pCO_2 on Earth may have been as great as 14 bars, although the results must be considered as only suggestive. One interesting finding was that Rayleigh scattering significantly increases the albedo of atmospheres high in CO_2. However, Durham & Chamberlain (1989) did not take into account CO_2 condensation, and the effect of the resulting high-albedo CO_2 clouds on the early climate of the planets (see below, Kasting 1991b).

EARLY HOT, H_2O-RICH ATMOSPHERES

In a series of papers, Matsui & Abe (1986a,b; Abe & Matsui 1988), and more recently Tajika & Matsui (1992), explored the question of the radiative effects of dense steam atmospheres that may have been present during accretion of the terrestrial planets, and subsequent atmospheric conditions. The source of the water would have been infalling planetesimals (inner Solar System objects and/or late accreting material from the outer Solar System) that released their volatiles when impacting at speeds exceeding 2 to 3 km/sec. Partial release of volatiles by impacting planetesimals may have begun when Venus and Earth were only ~1% of their present masses, and complete release would have occurred when the planets were ~10% of their final masses. For Mars, by contrast, devolatization of impactors would have begun later in the accretionary history, and a significant fraction of the volatiles that went into the formation of Mars was probably not released during accretion (Pollack 1991).

The water and CO_2 released during accretion may have stayed in the atmospheres of Earth and Venus leading to a runaway greenhouse powered by solar and accretionary heating (Matsui & Abe 1986a,b; Zahnle et al 1988). If accretion took ~10^8 years, then enough energy would have been released to create steam atmospheres. Eventually the surface temperatures could have reached the melting point temperatures of surface rocks (~1500 K) when solution of the water in surface melt rocks would have helped to stabilize the atmosphere. Subsequent to the steam atmosphere phase, stabilization of Earth's CO_2-rich atmosphere could have been established through the carbon cycle (Tajika & Matsui 1992).

TOO COLD: THE ICEHOUSE OF MARS

Mars is a frozen wasteland. The average temperature is 218 K, or −55°C (Table 2), and even with the large seasonal and diurnal fluctuations that occur, liquid water does not exist. The low atmospheric pressure on Mars (7 mb) further prevents the condensation of liquid water (Kahn 1985). The low surface temperatures on Mars also mean that there can be very little water vapor in the atmosphere. Appreciable amounts of water are apparently stored in permanent water ice caps and as permafrost beneath the surface (Pollack 1991). Geologic evidence has been interpreted as suggesting at least 440 m of water, present as ice, groundwater, and ground ice. (Overlying the water ice caps at both poles are "seasonal" ice caps of solid carbon dioxide which condense from the atmosphere in winter and sublime to the atmosphere in summer.) Given this deep freeze, water vapor feedback on Mars today is so weak that even with twenty times the surface pressure of CO_2 as Earth, Mars produces less than half the greenhouse warming, and most of the 33 degrees of greenhouse warming on Earth comes from water vapor.

Yet, considerable evidence exists for a warmer, wetter past for Mars. Images of the Martian surface returned by *Mariner* and *Viking* spacecraft show evidence of ancient riverbeds—channels cut into the surface presumably by running water—with dendritic drainage patterns, and meandering paths. Valley networks occur in the ancient cratered terrain of Mars, indicating liquid water on the surface of the planet ~3.8 billion years ago (McKay & Davis 1991). The valleys and channels may have emptied into standing bodies of water—large lakes [possibly ice-covered (McKay 1991)] or small oceans (Parker et al 1993, Baker et al 1991). Liquid water on the planet's surface at much later periods of time is also considered a possibility by some workers (McKay 1991). This evidence suggests that Mars was once warm enough to sustain liquid water ($T > 273$ K). The most plausible explanation of this prior warmth and greater atmospheric pressure was a strong greenhouse warming from an early, much denser CO_2 atmosphere.

Although impact degassing may have been ineffective on Mars, carbon dioxide and water vapor were probably outgassed early in the planet's history. The amount of outgassing, however, is uncertain. McElroy et al (1976, 1977) estimated 8 to 133 m of outgassed water. From evidence of volcanism, Greeley (1987) estimated a global layer 46 m deep released in the first 2 billion years.

There are some indications that very little water remains in the upper mantle of Mars: Dreibus & Wanke (1987) estimated 18–36 ppm in Mars' mantle, and SNC meteorites (apparently originating on Mars) have low water contents of 16–52 ppm.

Strangely enough, the presence of an early massive CO_2 atmosphere on Mars, permitting liquid water to occur on the surface, could be part of the reason Mars became such an inhospitable place. If Mars had a 1 bar CO_2 atmosphere, weathering of silicates could have entirely converted the CO_2 to carbonate rocks in ~10 million years (Pollack 1991). Without replenishment of CO_2, the warm, greenhouse conditions would have been lost quite quickly. On Earth, CO_2 precipitated as carbonate rock is recycled to the atmosphere by a combination of chemical reactions and plate tectonics (Figure 3). Replenishment of atmospheric CO_2 comes about through plate tectonics and resulting volcanism. The heating of carbonate rocks as ocean plates descend into the Earth's mantle causes decomposition of $CaCO_3$, and the resulting CO_2 is released to the atmosphere through volcanoes (Walker et al 1981). Mars shows no signs of a history of plate tectonic activity, but a kind of recycling might have taken place on early Mars, where large-scale lava flows could have buried carbonates and caused release of CO_2. Higher early heat flow (~3 to 5 times present values) might have made such a process possible (Pollack 1991).

Impact bombardment might have provided a more effective early source of heat on Mars (Carr 1989), although impact heating may not have occurred fast enough to prevent the loss of considerable atmospheric CO_2 through weathering reactions, and impact erosion may have actually removed volatiles from the Martian atmosphere (e.g. Ahrens 1993, Hunten 1993). Pollack (1991) went further in proposing a possible feedback in which the weathering rate would have adjusted to the resupply of CO_2, since the fraction of time liquid water was available (when ground temperatures hovered around 273 K) could have been controlled by the weathering rate needed to balance the CO_2 outgassing rate. This situation would have ended as the global heat flow decreased, bombardment slowed, and global volcanism died down. Carr noted that the early valley drainage networks approximately coincided with the time of heavy bombardment. The decrease in atmospheric pCO_2 to its low present value may have occurred as CO_2 was adsorbed onto regolith, or possibly by dry carbonate formation from CO_2 and water vapor (Pollack 1991). Carbonates have been detected on the surface of Mars by remote sensing (Pollack 1991), and in SNC meteorites of probable Martian origin.

However, all of these scenarios may be moot—a recent simulation of the climate of early Mars (Kasting 1991b) predicted that the low temperatures would permit the formation of CO_2 ice clouds in the upper troposphere. These clouds would further cool the surface by reducing the tropospheric lapse rate (because of the release of latent heat), and by reflecting additional sunlight back

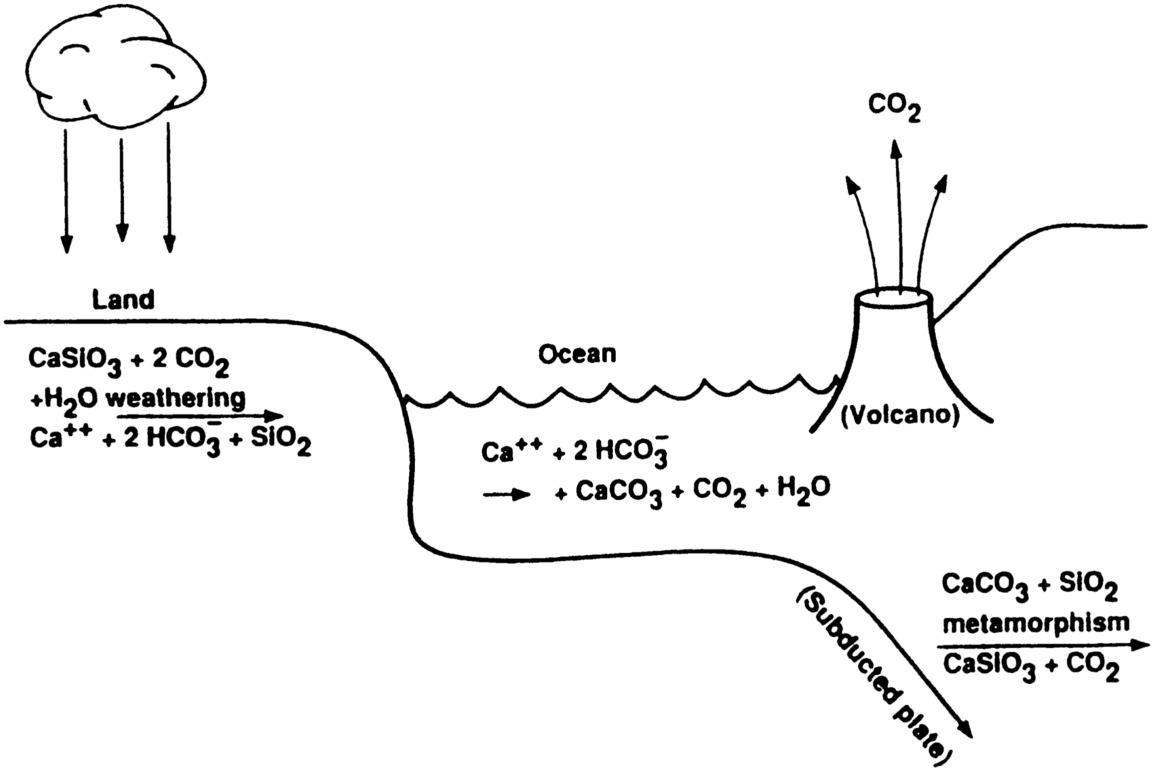


Figure 3 Schematic diagram of the biogeochemical cycle of carbon on Earth (courtesy of J. F. Kasting).

to space. Because CO_2 clouds are poor IR absorbers, their contribution to the greenhouse effect should be relatively small; their primary influence would be to cool the planet. Thus, Kasting's work suggests that a CO_2/H_2O greenhouse for early Mars does not work. This opens up the possibility that other greenhouse gases (CH_4, NH_3?) may have provided a warm climate.

Mars apparently represents a "runaway glaciation" model of planetary development (Hart 1978). Could this have happened to Earth? In early work, Schneider & Gal-Chen (1973) suggested that a decrease of 1.6% in the solar constant (equivalent to Earth being 0.8% farther from the Sun at present) would lead to runaway glaciation, whereas Wetherald & Manabe (1975) found that Global Climate Model (GCM) experiments would not produce a runaway glaciation on Earth even if the solar constant was decreased by 4%. In Hart's (1978) modeling effort, using an early atmosphere averaging ~1.2 atm, composed of CO_2 followed by increases in CH_4 and other reduced gases, if the Earth-Sun distance was increased by 0.01 AU, then Earth became terminally glaciated about 2 billion years ago. Clearly, however, the predictions for runaway glaciation of Earth depend upon assumptions about the composition and density of the early atmosphere.

In the history of Mars, we also see the effect that a planet's size can have on its evolutionary path. Information from the study of the Moon showed that the once volcanically active body was geologically dead by about 3 billion years ago. The reason is most likely that the Moon is too small a planetary body (equatorial diameter of 3,476 km; only ~1/4 the diameter of Earth, with about 1/80th of Earth's mass) to retain the internal heat trapped during accretion and generated by radioactive elements. Geologic activity, volcanism, and outgassing shut down after less than 1.5 billion years, and with only 1/6th the Earth's gravitational force, the Moon quickly lost to space whatever atmosphere it may have had.

The situation for Mars was apparently similar, but not so drastic. Mars is apparently too small (equatorial diameter = 6,794 km; ~1/2 Earth's diameter) to have retained its internal heat for the entire 4.6 billion year history of the Solar System, and geologic activity, carbon cycling (if it existed), and volcanism apparently wound down and ended some 0.5 to 1 billion years ago (Hansson 1991). As the Martian interior cooled, degassing of CO_2 would have slowed, and carbon dioxide would have been drawn out of the atmosphere through silicate-rock weathering reactions and converted to calcium carbonate rocks in the crust—a process that could continue only so long as the temperature remained above the freezing point of water. This scenario may explain why Mars' atmospheric pressure is so close to the triple point (the point where solid, liquid, and vapor phases of water coexist).

JUST RIGHT: MOTHER EARTH

Evidently, Earth averted both the deep freeze of Mars and the greenhouse hell of Venus. But how? Pollack (1991) asks the critical question, "Was it just blind luck that enabled the Earth's mean temperature to vary relatively little over almost its entire history so that oceans and life could have persisted over such an extended period, despite a varying solar luminosity?" The alternative to blind luck is that some important feedbacks exist that have controlled the Earth's climate history. Such feedback processes may be geochemical and/or biogeochemical.

Earth is clearly set apart from the other terrestrial planets by its wetness. The amount of liquid water present near or at the surface of Earth is 1.4×10^9 km^3, more than 97% of which is in the oceans (2% in polar ice and glaciers). This is equal to a layer of water 2.7 km thick covering the entire planet.

The interior of Earth also seems to contain significant amounts of water, although the exact amount is not well known. Estimates range from ~60 to > 200 ppm in the upper mantle (Dreibus & Wanke 1987, Jambon & Zimmerman 1990, Carr & Wanke 1992). Thus, Earth is apparently wet inside and out. Most of the Earth's water may have been derived from a late volatile-rich veneer, with any earlier water reacting with iron to form FeO and H_2, the latter of

1994ARA&A..32...83R

which would have been lost (Dreibus & Wanke 1987). The wet interior may be the result of melting of the Earth's surface during accretion, as a result of the development of a steam atmosphere, allowing impact devolatized water at the surface to dissolve into the molten rock (Abe & Matsui 1988). Because of Mars' smaller size and greater distance from the Sun, the Martian surface may not have melted to the same degree, preventing the uptake of water. Or possibly Mars acquired a late volatile-rich layer, which was not folded into the interior (as with the more geologically active Earth) and instead remained as a surface veneer (Carr & Wanke 1992).

An examination of the evolution of surface conditions on Earth quickly leads to what has been termed "the faint young Sun paradox": Calculations showed that the early Earth, with an atmosphere of current composition, would be too cold for liquid water, and hence life, as recently as 2 billion years ago; however, the geologic record clearly shows evidence for life as far back as 3.5 billion years ago, and liquid water back to ~3.8 billion years ago (Figure 4).

Sagan & Mullen (1972) suggested that reduced gases (mainly NH_3 and CH_4) in Earth's early atmosphere could have counteracted the faint young Sun problem by providing an additional greenhouse effect. In particular, NH_3 mixing ratios of only 10^{-5} to 10^{-4} were calculated to produce enough additional greenhouse heating to counteract the faint young Sun. However, these reduced gases

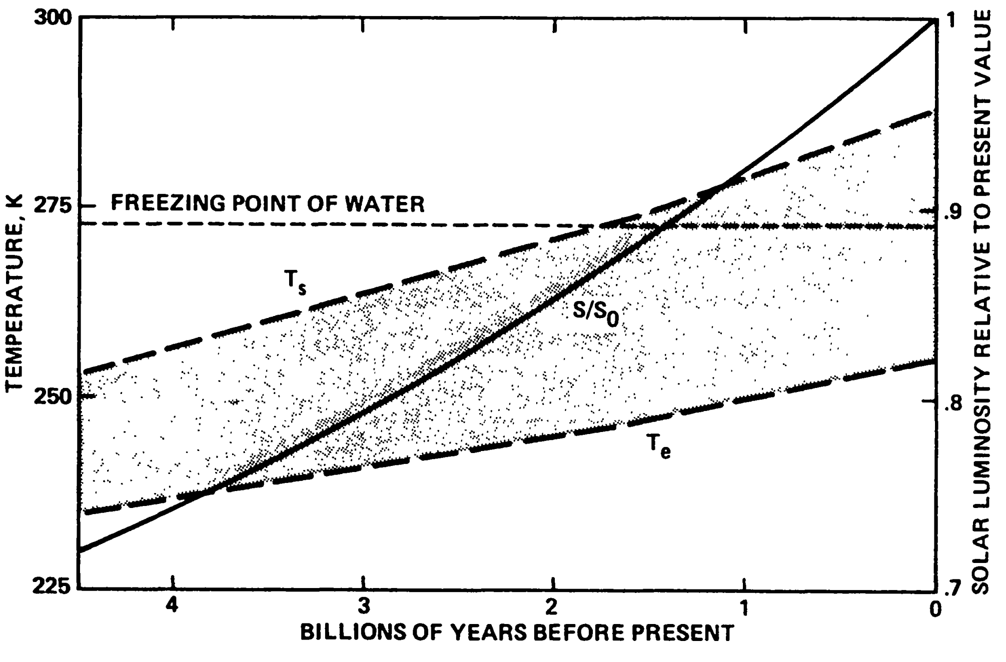


Figure 4 The faint young Sun problem as calculated with a one-dimensional, radiative/convective climate model. The solid curve is Gough's (1981) solar-luminosity parameterization. The dashed curves represent the effective radiating temperature T_e and the surface T_s. The shaded area shows the magnitude of the greenhouse effect. (After Kasting & Grinspoon 1991.)

were found to be too unstable with respect to photodissociation to provide a long-term source of greenhouse heating (Kuhn & Atreya 1979, Owen et al 1979). Owen et al (1979) suggested that a CO_2-rich early atmosphere could have provided the additional greenhouse warming to offset the faint young Sun problem.

The silicate/carbonate rock cycle is a geochemical conveyor belt driven by heat released by the mantle (Berner et al 1983). At present, it takes ~100 million years for new sea floor to upwell at mid-ocean ridges, spread to subduction zones, and, finally, to plunge into the mantle—long enough for atmospheric CO_2 to have cycled through the rock shell of the Earth many times over. Over the 4.6 billion years of planetary evolution there has been a gradual slowing of this conveyor belt, and probably a reduction in the amount of volcanic carbon dioxide released to the Earth's atmosphere.

However, although the greenhouse effect of an early atmosphere much richer in CO_2 is invoked to explain the warm conditions of the early Earth, this leads to the question of how the "thermostat" was regulated subsequently in the face of increasing solar luminosity. A stabilizing climatic feedback process is strongly indicated.

Walker et al (1981) were the first to propose that, on geological time scales, climate is stabilized by factors affecting the rate at which calcium silicate rocks are geochemically weathered. A key point is that the rate at which CO_2 is removed from the atmosphere by rock weathering increases as temperature increases. The two major reasons for this are that: 1. rainfall and runoff increase, exposing more bedrock to erosion; and 2. respiration of soil organisms increases. In the long run, a higher CO_2 level in soil increases weathering rates, and like increased weathering from increased rainfall, this is a negative feedback.

If, for some reason, surface temperature were to fall, the temperature-weathering feedback would cause the removal rate of CO_2 from the atmosphere to fall, while carbon dioxide continued to outgas. The net result would be more atmospheric carbon dioxide, and a greenhouse warming partly compensating for the initial cooling. Conversely, were surface temperature to rise, the increased weathering would drive atmospheric CO_2 levels down, creating a cooling opposed to the initial warming. The long-term effect of weathering is therefore to stabilize global temperature.

The gradual reduction in the CO_2 content of the atmosphere as Earth cooled was probably partly the effect of decreasing geologic activity as radioactive elements decreased and as Earth's internal heat dissipated. Pulses of activity seem to have occurred, as evidenced by variations in ocean-floor spreading rates (Larson 1991). There is also evidence from variations in carbon isotopes and organic-carbon burial rates that during the last 600 million years (the Phanerozoic Eon), atmospheric CO_2 rose and fell by significant factors (Berner 1993).

For example, ~100 million years ago (the mid-Cretaceous Period) was a time of great warmth. This period correlates with a marked increase in the rates of sea-floor creation, and therefore subduction and volcanism. Earth's climate was 10–15°C hotter and the poles probably ice-free, quite likely in response to greenhouse warming from an atmosphere 4–10 times richer in carbon dioxide (Caldeira & Rampino 1991).

Biogeochemical cycling of carbon dioxide helped Earth to avoid the greenhouse that entrapped Venus, but how did we escape becoming a frozen planet like Mars? Early work by Budyko (1969) and Sellers (1969) suggested that a 2 to 5% decrease in the solar constant would lead to an ice-covered Earth. In their models, a small reduction in solar luminosity resulted in the advance of ice and snow cover toward the equator; this additional ice and snow reflected more solar radiation back to space and further cooled the planet. If solar luminosity is reduced beyond some critical value in these models, this ice-albedo feedback results in the catastrophic, irreversible freezing of the entire Earth's surface—the "white Earth catastrophe."

The temperature dependence of the rate of CO_2 uptake by silicate weathering and the negative feedback that it creates may have stabilized Earth against the ice-albedo catastrophe (Walker et al 1981). Walker et al suggested that, given the low early solar luminosities, high atmospheric CO_2 concentrations would be necessary to produce sufficient CO_2 consumption by silicate weathering to balance sources of atmospheric CO_2. Berner, Lasaga & Garrels (1983) incorporated this feedback in a carbonate-silicate cycle model of the global carbon cycle.

However, work by Marshall et al (1988) suggested that the silicate-weathering feedback could produce cold global climates if the continents were clustered near the equator. Exposed silicate rock would then be bathed in a relatively warm and wet environment, conducive to CO_2 consumption by weathering. Thus, for example, lower solar luminosity and the presence of an equatorial, Late Proterozoic (1 billion to 600 million years ago) supercontinent may both help to explain the widespread low-latitude glaciation (Marshall et al 1988, Hambrey & Harland 1985, Walter 1979).

The calculations of Marshall et al (1988) suggested that the silicate-weathering/CO_2 feedback could stabilize the global ice line at low latitudes. However, reanalysis of their model has indicated that this is not possible (Caldeira & Kasting 1992a). Because atmospheric pCO_2 responds much more slowly ($\sim 10^5$ yr) than does sea ice and snow cover (< 1 yr), the negative feedback proposed by Marshall et al (1988) does not apply to rapid fluctuations in the ice line. The silicate-weathering feedback could not act rapidly enough to buffer the Earth against a catastrophic ice advance. Indeed, during the Pleistocene (the last 2 million years) the correlation between glacial ice mass and atmospheric pCO_2 is negative, not positive (Barnola et al 1987). A perturbation analysis (Cahalan

& North 1979) of the Marshall et al (1988) energy-balance model shows that there is no stable ice line equatorward of about 30° latitude, regardless of atmospheric pCO_2 level. Hence, in their model, a low-latitude glaciation should run away to the globally ice-covered state. Could this have happened during the Late Proterozoic?

To study this question, Caldeira & Kasting (1992a) developed a zonally averaged energy-balance climate model based on a one-dimensional radiative-convective climate model (Kasting & Ackerman 1986). At today's solar flux and CO_2 level, the model exhibits four possible steady states (Figure 5): 1. ice free ($x_s = 1$), 2. stable partial ice cover ($x_s = 0.95$), 3. unstable partial ice cover ($x_s = 0.28$), and 4. ice covered ($x_s = 0$).

If Earth was initially in the stable, partially ice-covered state, and was then subjected to a rapid perturbation (relative to the 10^5 yr response time of the silicate weathering feedback) that would either temporarily lower the effective solar flux to about 0.9, or produce transient glaciation equatorward of $x_s = 0.28$, the planet would fall into the ice-covered state ($x_s = 0$). Without any change in atmospheric pCO_2, an increase in solar flux by about 27% would be needed to melt the equatorial ice (Figure 5). At solar fluxes higher than this value, a reverse ice-albedo feedback would apparently completely deglaciate Earth.

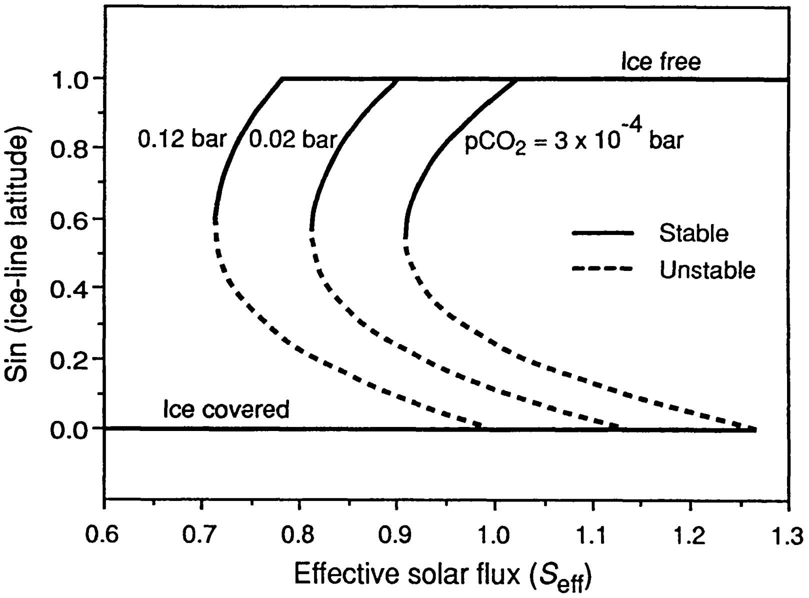


Figure 5 Steady-state ice lines (x_s) as a function of effective solar luminosity, S_{eff}, for three values of atmospheric pCO_2. If a perturbation were to shift Earth from its present state ($S_{eff} = 1$, $pCO_2 = 3 \times 10^{-4}$ bar, $x_s = 0.95$) to an ice-covered state today ($S_{eff} = 1$, $pCO_2 = 3 \times 10^{-4}$ bar, $x_s = 0$), then sufficient CO_2 (~0.12 bar) would accumulate in the atmosphere within ~30 million years to make the ice-covered state unstable. The model would then shift to the ice-free state ($S_{eff} = 1$, $pCO_2 = 0.12$ bar, $x_s = 1$) and silicate-rock weathering would begin to remove the excess CO_2 from the atmosphere. (After Caldeira & Kasting 1992a.)

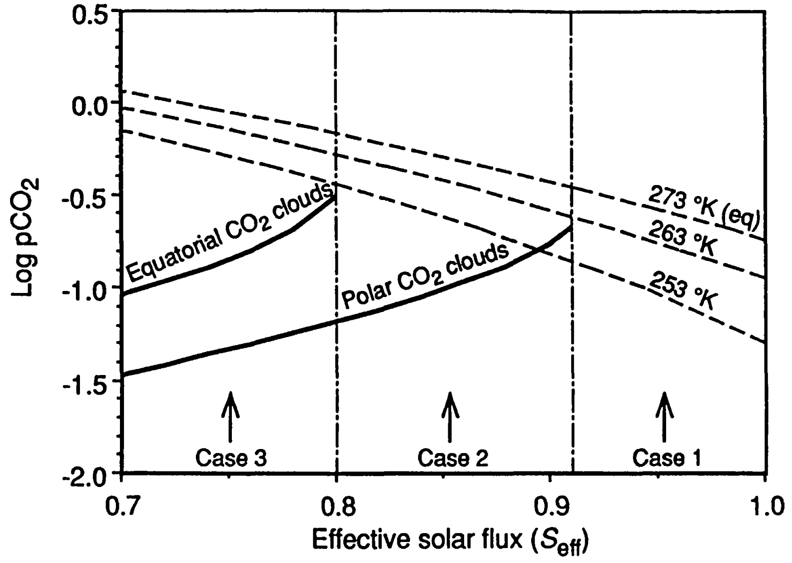


Figure 6 Solid curves indicate the onset of polar and global CO_2-cloud cover; dashed curves show equatorial surface temperature as a function of effective solar luminosity and atmospheric pCO_2. Case 1: Starting from an ice-covered state with low atmospheric pCO_2, volcanic CO_2 would accumulate in the atmosphere and initiate ice-melting prior to the formation of CO_2 clouds. Case 2: Polar CO_2 clouds would form prior to the onset of equatorial ice-melting. Case 3: CO_2 clouds would form globally prior to the onset of ice-melting, in which case CO_2-global warming would likely be incapable of melting the ice. (After Caldeira & Kasting 1992a.)

But this is not what would actually happen were the present Earth to freeze. In the ice-covered state, little or no silicate rock would be exposed to weathering, so CO_2 from metamorphic and mantle sources could accumulate in the atmosphere at a rate of about 8×10^{12} mol yr^{-1} (Holland 1978). In less than ~30 million years, atmospheric pCO_2 would build up to nearly 0.12 bar, and equatorial ice would become unstable. This evolution corresponds to Case 1 in Figure 6.

However, if the change from stable partial ice cover to global glaciation occurred earlier in Earth's history, when solar luminosity was lower, the results might then be quite different. The reason is the early formation of clouds of CO_2 ice, wrapping the planet in a high-albedo blanket (Kasting 1991b). The points where CO_2 clouds begin to form in the Caldeira & Kasting model are indicated by the solid curves in Figure 6. The earlier phase of Earth's history is divided into two parts, labeled "Case 2" and "Case 3." When S_{eff} is less than about 0.92, i.e. earlier than about 1 billion years ago (Gough 1981), the model predicts that CO_2 clouds would start to form at the poles (Case 2). When S_{eff} is less than about 0.8, i.e. earlier than about 3 billion years ago (Gough 1981), CO_2 cloud cover would extend all the way to the equator.

If the albedo of such a CO_2 cloud-covered planet were as high as 0.8, the model predicts that Earth would not be able to emerge from that state, even at the present solar luminosity. Thus, if Earth had experienced runaway glaciation prior to about 3 billion years ago, the situation might have been effectively irreversible. These considerations imply that our planet might now be uninhabitable had it not been warm during the early part of its history.

THE GAIA HYPOTHESIS

The most controversial idea related to the Goldilocks Problem may be the Gaia Hypothesis (Lovelock & Margulis 1974; Lovelock 1979, 1989), which holds that living organisms on Earth actively regulate atmospheric composition and climate in the face of challenges such as the increasing luminosity of the Sun. Versions of the Gaia Hypothesis range from "strong Gaia," which proposes that life maintains planetary conditions at some "optimum" for living things, to "weak Gaia," which maintains that biological processes and feedbacks affect global climate (see Schneider & Boston 1991). Although a compelling case has by no means been made for the "strong"—planetary homeostasis—version of Gaia, most earth scientists would accept a "weak" version—that biological processes and feedbacks are important (see Schneider & Boston 1991). It is worth noting that Krumbein & Schellnhuber (1990) estimated that the total mass of all organisms that have ever lived on Earth as ranging from 2.5×10^{26} to 2.5×10^{33} g, or from 10% to 10^5 times the mass of Earth itself. This in itself suggests that the biosphere cannot be ignored as a major player in cycling energy and matter during the history of the planet.

Although many geochemists believe carbon dioxide variations over geologic time can be explained primarily through abiotic interactions such as rock weathering (e.g. Holland et al 1986), such explanations are increasingly challenged by advocates of Gaia, or Gaian-type feedback mechanisms (Lovelock 1979, 1989; see also Schneider & Boston 1991). For example, the weathering rate feedback, in which CO_2 is removed as it builds up and causes planetary warming, is now understood to be accelerated by organisms (Volk 1987). Important changes apparently took place $\sim$3 billion years ago when the first soil-forming microorganisms created an environment in which CO_2 concentrations in contact with weathering rocks could be greatly elevated (Schwartzman & Volk 1989), and $\sim$100 million years ago when angiosperms replaced gymnosperms as the dominant plant group, possibly causing an increase in weathering rates through higher rates of root respiration and increased levels of soil pCO_2 (Volk 1989a).

At the same time, other forms of life may be accelerating the degassing of CO_2 from carbonate rock, and, hence, increasing the CO_2 supply to the atmosphere. For example, prior to the Cretaceous ($\sim$140 million years ago), most carbonate accumulated in shallow-water environments and open ocean carbonate accumulation was a minor component in the global carbon cycle (Boss & Wilkinson 1991). The evolution of calcareous plankton and their spread in the Cretaceous changed that situation. Since that time, there has been a gradual shift from shallow-water to deep-water carbonate accumulation (Opdyke & Wilkinson 1988). Deep-water carbonate is transported rapidly to subduction zones; hence, there may already be enhanced CO_2 degassing due to the subduction-zone metamorphism of the tiny shells of open ocean organisms (Volk 1989b; Caldeira 1991, 1992).

Other Gaian mechanisms that have been invoked to produce a planetary homeostasis include the emission and uptake of greenhouse gases such as CO_2 and methane by life in the ocean and in soils, and the release of dimethyl sulfide gas (DMS), which provides cloud condensation nuclei for clouds (Charlson et al 1987), by ocean plankton. An increase in cloud condensation nuclei can cause clouds to become more reflective, thus increasing the planetary albedo and cooling the Earth. Cloud effects on solar UV have also been noted. Mass extinctions of ocean plankton, such as occurred at major geological boundaries, should decrease DMS and cloud condensation nuclei, causing drastic climate warming (Rampino & Volk 1987), and some evidence of this has been found in the record.

Gaian mechanisms might go further and affect the inner workings of Earth, and thus impact the carbon cycle by changing the major driving force behind volcanic outgassing and the return of volatiles to the mantle (so-called ingassing). For example, an increase in volatiles (H_2O, CO_2) in the upper mantle lowers the kinematic viscosity, allowing more rapid convection (McGovern & Schubert 1989). Periods when volatiles were more efficiently ingassed into the mantle should be followed by increased convection, and presumably faster sea-floor spreading rates. Such a situation may have arisen in the Cretaceous Period of Earth history (140 to 65 million years ago). As we mentioned, the evolution and diversification of calcareous plankton in the Early Cretaceous caused a significant increase in carbonate deposited on ocean crust. This carbonate was rapidly recycled by subduction, with a fraction going back into the mantle as CO_2 (in addition to water from hydrated ocean crust and sediments).

Such an increase in mantle volatiles is predicted to decrease mantle viscosity, possibly leading to a period of increased spreading rates on ocean ridges, increased subduction, and volcanic outgassing of CO_2. Mid-Cretaceous (120 to 80 million years ago) spreading rates were exceptionally high (Larson 1991). The increased convection would act to cool the mantle, causing an increase in viscosity, and hence a slowdown in convection and sea-floor spreading. The pulse of activity could be self-limiting, and might run in cycles tens to hundreds of millions of years long.

Another possibility is that since the viscosity of the mantle in the vicinity of the subduction zones would be most dramatically decreased with the arrival of the subducted carbonate, the subducting slabs may begin to sink more rapidly. If slab pull is the determining force for plate motion (Kearey & Vine 1990), ocean crust production rates might increase quite rapidly. Thus, strangely enough, the evolution of calcareous plankton may have led to increases in ocean crustal production rates.

CONTINUOUSLY HABITABLE ZONES

The habitable zone (HZ) around stars has been defined as that region in which planetary temperatures are neither too high nor too low for life to develop, whereas the continuously habitable zone (CHZ) around a main sequence star is defined as the region around the star within which planetary temperatures remain within the temperature constraints for habitability, taking into consideration the evolution of the star's luminosity (Hart 1978, 1979). (Others have defined habitability in human terms, suggesting that habitable planets are those on which large numbers of people can live comfortably, e.g. Dole & Asimov 1964). Using computer simulations of Earth's atmospheric composition and surface temperatures, Hart (1978, 1979) considered solar luminosity changes, variations in Earth's albedo, atmospheric pressure and composition, the greenhouse effect, variation in biomass, and a variety of geochemical processes, in order to calculate CHZs around main sequence stars.

Hart's finding of a relatively narrow CHZ around the Sun (0.95 AU to 1.01 AU) during its lifetime, and even narrower CHZs around other stars, prompted responses from planetary scientists and climate modelers (Owen et al 1979, Schneider & Thompson 1980). Hart seemed to be reducing an important variable in the Drake Equation to a very low number, and hence his results had important repercussions for the fledgling field of SETI (Search for Extraterrestrial Intelligence). Hart's modeling has been criticized on the basis of a number of his assumptions that govern the sensitivity of his climate model (e.g. the composition of the early atmosphere, that the fraction of the planet covered by clouds is proportional to the total mass of water vapor in the atmosphere, and that a strong negative feedback exists between the amount of clouds and surface temperatures), as well as for his particular scenarios of planetary and atmospheric evolution.

Recently, the question of habitable zones around main sequence stars has been taken up by Kasting et al (1993). The inner edge of the HZ is determined in their model by the loss of water through photolysis and hydrogen escape. The outer edge of the HZ is determined by the formation of CO_2 clouds that cool a planet's surface by increasing its albedo and by lowering the convective lapse rate. Conservative estimates in the Solar System for these distances are 0.95 and 1.37 AU. The width of the HZ is slightly greater for planets that are larger than Earth and which have higher N_2 partial pressures in their atmospheres. Climate stability between these two limits is controlled by the weathering rate/temperature feedback mechanism, in which higher temperatures lead to higher weathering rates of silicate rock, which lowers atmospheric CO_2 and hence lowers surface temperature and weathering rate, and vice versa.

During the evolution of the Sun, the HZ evolves outward (Figure 7) as the Sun's luminosity increases. Kasting et al (1993) conservatively estimated that the CHZ over 4.6 billion years ranges from 0.95 to 1.15 AU (without cold starts—that is, a planet that starts out cold, with clouds of frozen CO_2, probably cannot escape global glaciation), or a width of 0.2 AU, which is much greater than Hart's (1978 and 1979) estimates of 0.06 and 0.046 AU, respectively.

Kasting et al (1993) have looked at CHZs around stars from 0.5 to $1.5M_\odot$ (stars more massive than $1.5M_\odot$ have lifetimes too short to be interesting from a habitability standpoint), and found that all of the calculated CHZs are 4 to 20 times wider than those calculated by Hart (1979). Planets around late K and M type stars may not be habitable as they can become trapped in synchronous rotation as a result of tidal damping, although mid to early K stars provide another possible location for habitable planets in addition to G stars like the Sun. This has important ramifications for SETI.

THE FATE OF THE EARTH

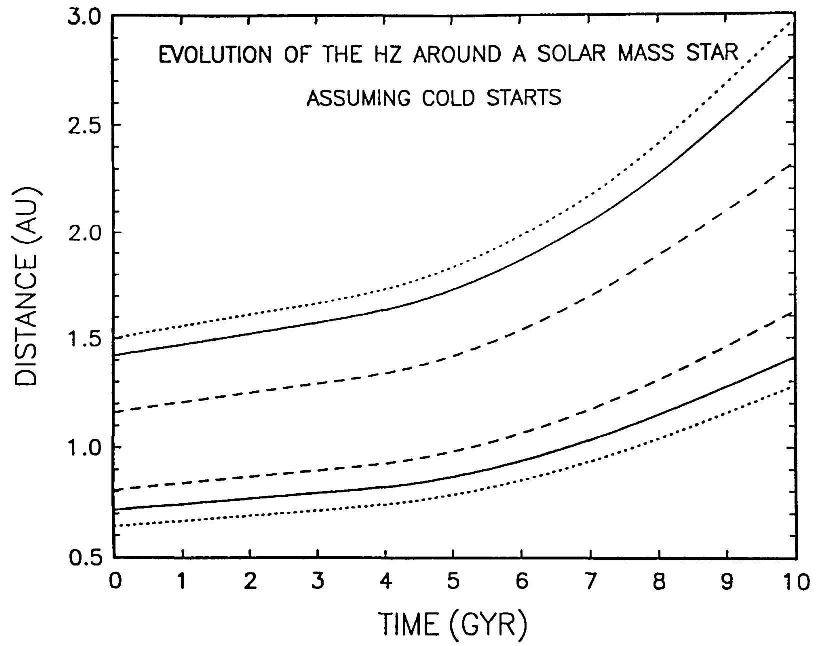
When will solar luminosity be so high that liquid water can no longer exist on Earth's surface? To answer this question, some estimate needs to be made regarding the future atmospheric content of greenhouse gases. Lovelock & Whitfield (1982) proposed that atmospheric CO_2 content will tend toward zero in approximately 10^8 yr from now, due to a CO_2-weathering feedback of the type proposed by Walker, Hays & Kasting (1981).

Photosynthetic organisms, at the base of the food chain, extract carbon from the atmosphere and from dissolved CO_2. These carbon reducers, dependent on an adequate atmospheric pCO_2 for their survival, are the source of carbon for the rest of the biosphere. Hence, atmospheric pCO_2 reductions brought about by increased solar luminosity could effectively cut off the carbon flux to the biosphere. At present-day geologic CO_2 degassing rates, this could happen in less than 1 billion years, unless organisms evolve to more efficiently exchange CO_2 with the atmosphere (Caldeira & Kasting 1992b).

The shift of carbonate deposition to the deep oceans with the appearance of calcareous plankton about 140 million years ago has already been mentioned. If subducting deep-ocean carbonate efficiently degasses its CO_2 to the atmosphere or ocean, then a complete transfer of the sedimentary carbonate mass to the pelagic realm could increase the CO_2 supply and chemical weathering rates by an order of magnitude. This might mean that thermal limits and/or loss of water, and not CO_2-starvation may ultimately limit the lifespan of the biosphere.

However, even without this biologically mediated accelerated CO_2-degassing, atmospheric CO_2 content would probably not go all the way to zero. Model cal-

1994ARA&A..32...83R



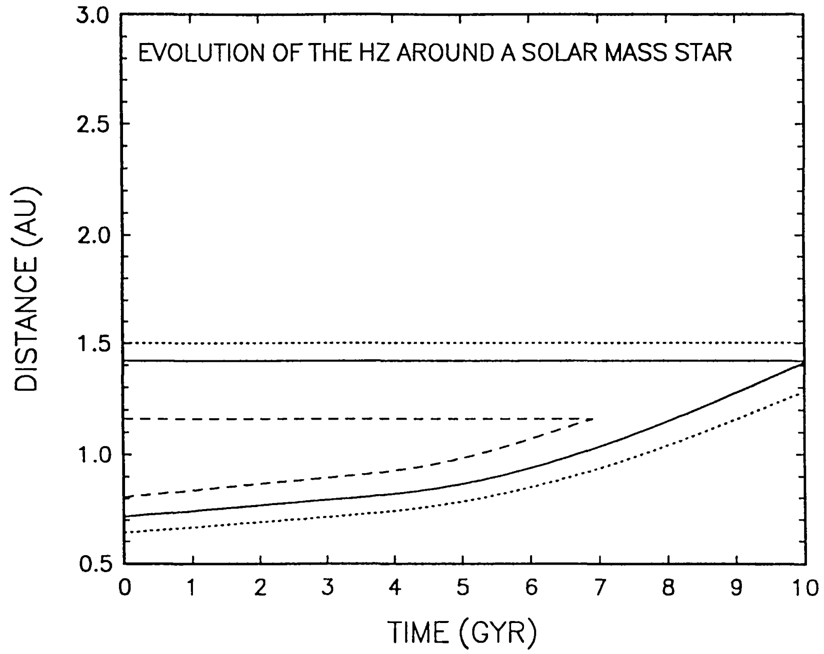


Figure 7 Evolution of habitable zone around a 1 $M_\odot$ star for two different assumptions concerning the possibility of "cold starts": (*top*) cold starts permitted, (*bottom*) no cold starts allowed. The term cold start refers to whether or not a planet will warm up once the stellar luminosity increases to the appropriate critical values (see text). The three pairs of curves correspond to the different habitability estimates discussed in the text: long dashes—water loss and first CO_2 cloud condensation limits (most conservative); solid curves—runaway and maximum greenhouse limits; dotted curves—recent Venus and early Mars limits (most optimistic). (After Kasting et al 1993.)

culations by Caldeira & Kasting (1992b) indicated that when the Earth's mean temperature becomes warm (> 300 K), silicate dissolution may proceed more rapidly than CO_2 and H_2SO_4 are supplied to the atmosphere by volcanism. This would leave an excess of Ca^{2+} and Mg^{2+} cations that could not be precipitated as carbonates or sulfates, which would probably precipitate in the oceans in a variety of silicate phases. Geologic carbon inputs to the ocean and atmosphere must be largely balanced by carbonate sedimentation, suggesting that the oceans would continue to be saturated with respect to some carbonate phase.

If future ocean waters are in approximate chemical equilibrium with silicates and carbonates, ocean pH would be buffered close to its present value. If ocean pH increases by no more than half a pH unit and the oceanic Ca^{2+} concentration changes by no more than an order of magnitude, then application of carbonate equilibria indicates that atmospheric pCO_2 would be reduced no more than two orders of magnitude from today's value. Even if ocean chemistry is not this well buffered, Henry's law implies that more CO_2 will be partitioned into the atmosphere on a warmer Earth, so the atmosphere may still contain up to a few ppm of CO_2.

According to the calculations, if the atmospheric pCO_2 is taken to be 1 ppm in the distant future, Earth's mean global surface temperature approaches 80°C in about 1.5 billion years. If there is more CO_2 in the atmosphere, this temperature could be reached sooner. As the surface temperature approaches $\sim$80°C, the stratospheric H_2O mixing ratio reaches $\sim$2.5%. Above this mixing ratio, the loss of hydrogen to space is limited by the solar EUV heating rate to $\sim 6 \times 10^{11}$ atoms H cm^{-2} s^{-1} (Watson, Donahue & Walker 1981), giving a time scale for ocean loss of $\sim$1 billion years. Hence, liquid water on Earth could be completely eliminated $\sim$2.5 billion years from now. This would bring terrestrial life to its final close. After Earth's water is lost, silicate-rock weathering will cease; hence, volcanic CO_2 should accumulate in the atmosphere, creating a climate much like that of Venus.

ANCIENT GLACIAL DEPOSITS AND EARLY CLIMATE

Much of the climate and carbon-cycle modeling related to the long-term evolution of the Earth's climate uses paleoclimate information as constraints (e.g. Kasting 1989). A major constraint has been the occurrence of ancient glaciations, including an Early Proterozoic (Huronian) Glaciation at $\sim$2.3 billion years ago, and the widespread Late Proterozoic glaciations between about 1 billion years and 600 million years ago (Hambrey & Harland 1981). During glacial epochs, T_s for Earth could not have been more than 20°C (293 K) (Kasting 1989). However, these glacial periods occur at times when many models predict high atmospheric pCO_2, and thus warm planetary temperatures (Kasting 1989). The Huronian glaciation has been a particular problem, as most models of atmospheric composition at 2.3 billion years ago predict large amounts of atmospheric CO_2, making such an early glacial epoch difficult to produce (Kasting 1989). Evidence for the Late Proterozoic glaciations occurs in many parts of the world, some at sites that are reconstructed at low latitudes, suggesting a worldwide glaciation (Hambrey & Harland 1981).

All of this, however, rests on the correct identification of these periods as times of glaciation. Recent studies suggest that some of the sedimentary deposits that represent the primary evidence for glaciation may have been misidentified (Rampino 1992, 1994; Oberbeck et al 1993). The evidence of past glaciations is primarily in the form of so-called tillites—typically poorly sorted sedimentary rocks, with boulders in a fine-grained matrix, interpreted as analogues of the glacial deposits (till) of the more recent ice ages. Diagnostic criteria include faceted and striated stones, striated rock pavements, and laminated siltstones containing large "ice-rafted" clasts. Late Proterozoic tillites in Australia are up to 6,000 m thick (Hambrey & Harland 1981).

The first clue that something might be amiss was the fact that essentially all studies based on modern sedimentological and stratigraphic analysis have concluded that many recognized "tillites" were not deposited passively by melting glaciers, but are in reality debris-flow and related mass-flow deposits, with a significant component from fallout of coarse material [some deposits have been called "debris-rain" sediments (Rampino 1994)].

Based on these analyses, tillites have generally been reinterpreted as debris-flow deposits formed where glaciers deposited material into the sea or large lakes. The current models of sedimentation envisions these debris-flow sequences as generated directly by rapid dumping of debris at a marine ice front, or by deposition of unstable sedimentary accumulations on the shallow sea floor by iceberg rainout, with later re-sedimentation in submarine debris flows. Melting icebergs are inferred to have provided additional rainout of coarse material.

Re-examination of the characteristics of these deposits, however, shows that they apparently are similar in distribution and textures to the expected deposits of ballistic ejecta of large impacts (Oberbeck et al 1993, Rampino 1994). In ballistic sedimentation, debris ejected at high velocity from an impact site, and material produced by spall from the ground as the shock waves move outward, sweep rapidly away from the impact site(s). This ballistic debris strikes the ground at greater and greater velocity farther away from the impact, imparting an outward momentum to a mixture of target debris and entrained local material, which moves rapidly outward as a ground-hugging debris flow. Local sediments are entrained within the fast-moving flows. For a 100-km crater on the moon (with no atmosphere) the horizontal component of velocity of the debris flows could be as great as $\sim 300 \text{ m s}^{-1}$. On Earth, the largest ejecta would not be significantly affected by atmospheric drag. Known ejecta deposits on Earth are composed of poorly sorted debris containing abundant striated and faceted stones, resting in some places, on eroded and striated bedrock surfaces (Rampino 1994).

A number of climatic enigmas, such as the seeming conflict between an early CO_2-rich atmosphere and the Huronian glaciation, and the apparent worldwide extent of Late Proterozoic glaciations, might be solved if the tillite deposits associated with these ice ages, or at least some of them, are in reality impact debris (Oberbeck et al 1993, Rampino 1994).

TERRAFORMING

Terraforming, or planetary engineering—the transformation of non-habitable planets into habitable places—has received considerable scientific attention in recent years (e.g. Fogg 1989, McKay & Haynes 1990, McKay et al 1991). If Earth eventually becomes uninhabitable, then terraforming other planets may represent an alternative for life. Mars is the most likely candidate for terraforming, and a number of possible schemes for re-engineering that planet's atmosphere and climate have been suggested (McKay et al 1991). An early NASA study (prior to analysis of *Viking* mission data) suggested that the greenhouse warming potential of the CO_2 in the Martian polar caps could raise planetary temperatures above freezing (Averner & MacElroy 1976). It was proposed, for example, that a reduction in the albedo of the polar regions—by covering them with low-albedo dust or dark plants—could release significant amounts of CO_2 into the atmosphere. However, we now know that the NASA terraformers overestimated the CO_2 mass of these caps available for greenhouse warming—the "dry ice" caps at Mars' poles are almost entirely volatilized and recondensed each Martian year (Tillman et al 1993).

McKay et al (1991) have made the most extensive study of terraforming Mars. They suggest that if a large amount of volatiles exist on Mars in subsurface reservoirs, planetary engineering schemes could exploit runaway effects involving release of water and CO_2 from the polar caps and regolith. McKay et al (1991) estimated that an amount of CO_2 adsorbed onto surface rocks equivalent to $\sim$300 mbar could be released into the Martian atmosphere. However, the amount of carbon dioxide that exists either as adsorbed CO_2, or in carbonate rocks is poorly known, and further exploration of Mars is required to provide additional information on global inventories of volatiles such as CO_2, H_2O, and N_2.

In a speculative study, Lovelock & Allaby (1984) suggested the creation of a more habitable Mars by using chlorofluorocarbons as greenhouse gases. In their scenario, CFCs could be sent from Earth in missiles. However, the authors may have significantly underestimated the mass of CFCs needed, and a better approach might be to manufacture on Mars new greenhouse molecules engineered to strongly and broadly absorb infrared radiation at very low concentrations (McKay et al 1991).

With a long-term commitment, terraforming of planetary atmospheres could become a future goal of humanity (e.g. Savage 1993). As McKay et al (1991) note "If ... investigations indicate that it is feasible to make Mars habitable, the motivation to do so may depend on the potential for life (human and non-human) on Mars, the economic payoff, and the nature of other large-scale efforts that provide habitats in space."

CONCLUSIONS

The Goldilocks Problem touches on many of the latest developments relating to the evolution of planets and their atmospheres. Recent work has stressed the early evolution of the atmospheres of Venus, Earth, and Mars, and the light that is shed on the probability of habitable planets around main sequence stars.

The two key factors relating to the habitability of a planet seem to be the size of the planet and its distance from the Sun, although the conclusions are not straightforward. Venus is too close to the Sun. The planet underwent a runaway greenhouse, or moist greenhouse, early in its history. In the runaway greenhouse scenario, the relatively high temperatures on early Venus would have prevented water from condensing, creating a dense H_2O atmosphere and intense greenhouse effect. The lack of liquid water prevented the silicate-rock weathering reactions that act to remove CO_2 gas from the atmosphere and store it as solid calcium carbonate on the planet's surface, allowing CO_2 to build up to present high levels. Water was lost as the high Venus temperatures allowed H_2O vapor to reach high altitudes where the H_2O molecules were broken by solar UV, and escaped from the atmosphere driven by solar extreme UV radiation, sealing Venus' fate.

In the moist greenhouse scenario, liquid water was able to condense on early Venus, but the low lapse rate of the warm Venus atmosphere allowed water vapor mixing ratios to remain high at great altitudes, facilitating the breakup and escape of H_2O by solar UV. As in the runaway greenhouse model, the loss of water allowed the buildup of the present massive CO_2-rich Venus atmosphere.

Mars is too small. Although apparently experiencing warm temperatures and the presence of liquid water early in its history, as the small planet lost its internal heat, its internal geologic engines shut down. Mars lost the ability (if it ever had it) to recycle carbon and release volcanic CO_2 to provide a replenishment of atmospheric CO_2, and thus maintain a significant greenhouse effect. The planet became a frozen world.

Earth was apparently lucky, but more than blind luck was involved. The relatively large size of the planet has kept its internal heat from leaking away too rapidly, and plate-tectonic activity has maintained the recycling of carbon dioxide. Earth is far enough away from the Sun so that early temperatures, probably maintained by a CO_2-rich atmosphere, remained within the range of liquid water. The water-mediated silicate rock weathering reactions have been effective in removing CO_2 from the atmosphere, with a built-in temperature-dependence feedback, so that as the Sun's luminosity increased, the resulting warmer surface temperatures on Earth caused weathering rates to increase, thereby removing more of the atmospheric CO_2, and cooling the planet's surface. Thus, a balance has been maintained, probably with the help of additional feedbacks related to life. However, as the Sun's luminosity continues to increase, biological and abiological mechanisms for homeostasis may fail, and recent model calculations suggest a hot, dry, uninhabitable Earth within about a billion years.

Recent estimates of the continuously habitable zones around stars like the Sun suggest a width of ~0.2 AU (from 0.95 to 1.15 AU), considerably more optimistic than previous estimates of a continuously habitable zone only about 25% of that width. This has positive implications for the distribution of life in the Cosmos, and the search for extraterrestrial intelligence.

ACKNOWLEDGMENTS

We thank M. I. Hoffert and T. Volk for discussions and information, and J. F. Kasting for a critical review.

Literature Cited

Abe Y, Matsui T. 1988. *J. Atmos. Sci.* 45:3081–101

Ahrens TJ. 1993. *Annu. Rev. Earth Planet. Sci.* 21:525–55

Averner MM, MacElroy RR, eds. 1976. *On the Habitability of Mars. NASA SP414.* Washington DC:US Govt. Print. Off.

Baker VR, Strom RG, Gulick VC, Kargel JS, Komatsu G, Kale VS. 1991. *Nature* 352:589–94

Barnola JM, Raynaud D, Korotkevich YS, Lorius C. 1987. *Nature* 329:408–14

Berner RA. 1993. *Science* 261:68–70

Berner RA, Lasaga AC, Garrels RM. 1983. *Am. J. Sci.* 283:641–683

Boss SK, Wilkinson BH. 1991. *J. Geol.* 99:497–513

Budyko MI. 1969. *Tellus* 21:611–19

Cahalan RF, North GR. 1979. *J. Atmos. Sci.* 36:1178–88

Caldeira K. 1991. *Geology* 19:204–6

Caldeira K. 1992. *Nature* 357:578–81

Caldeira K, Kasting JF. 1992a. *Nature* 359:226–28

Caldeira K, Kasting JF. 1992b. *Nature* 360:721–23

Caldeira K, Rampino MR. 1991. *Geophys. Res. Lett.* 18:987–90

Carr MH. 1989. *Icarus* 79:311–27

Carr MH, Wanke H. 1992. *Icarus* 98:61–71

Charlson RJ, Lovelock J, Andreae MO, Warren SG. 1987. *Nature* 326:655–61

Chyba CF. 1990. *Nature* 343:129–33

Dayhoff MO, Eck R, Lippincott ER, Sagan C. 1967. *Science* 155:556–57

DeBergh C, Bezard B, Owen T, Crisp D, Maillard J-P, Lutz BL. 1991. *Science* 251:547–49

Dole SH, Asimov I. 1964. *Planets for Man.* New York:Random House

Donahue TM, Hartle RE. 1992. *Geophys. Res. Lett.* 19:2449–52

Donahue TM, Hodges RR Jr. 1992. *J. Geophys. Res.* 97:6083–90

Donahue TM, Hoffman JH, Hodges RR Jr, Watson AJ. 1982. *Science* 216:630–33

Dreibus G, Wanke H. 1987. *Icarus* 71:225–40

Durham R, Chamberlain JW. 1989. *Icarus* 77:59–66

Eberhardt P, Hodges RR, Krankowsky D, Berthelier JJ, Schultz W, et al. 1987. *Lunar Planet. Sci.* XVIII:252–53

Fogg MJ. 1989. *J. Brit. Interplanet. Soc.* 42:577–82

Gilliland RL. 1989. *Palaeogeogr. Palaeoclimatol. Palaeoecol. (Global and Planet. Change Sect.)* 75:35–55

Gough DO. 1981. *Sol. Phys.* 74:21–34

Gold T. 1964. In *The Origin and Evolution of Atmospheres and Oceans,* ed. PJ Brancazio, AGW Cameron, pp. 249–56. New York:Wiley

Goody RM, Walker JCG. 1972. *Atmospheres.* Englewood Cliffs, NJ:Prentice-Hall

Greeley R. 1987. *Science* 236:1653–54

Grinspoon DH. 1987. *Science* 238:1702–4

Grinspoon DH. 1993. *Nature* 363:428–31

Grinspoon DH, Lewis JS. 1988. *Icarus* 74:430–36

Gurwell MA, Yung YL. 1992. *Planet. Space Sci.* 41:91–104

Hambrey MJ, Harland WB, eds. 1981. *Earth's Pre-Pleistocene Glacial Record.* Cambridge:Cambridge Univ. Press

Hambrey MJ, Harland WB. 1985. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 51:255–72

Hansson A. 1991. *Mars and the Development of Life.* New York:Ellis Horwood

Hart MH. 1978. *Icarus* 33:23–39

Hart MH. 1979. *Icarus* 37:351–57

Head JW, Crumpler LS, Aubele JC, Guest JE, Saunders RS. 1992. *J. Geophys. Res.* 97:13,153–97

Henderson-Sellers A. 1978. *The Origin and Evolution of Planetary Atmospheres.* London:Adam Hilger

Henderson-Sellers A, Cogley JG. 1982. *Nature* 298:832–35

Holland HD. 1978. *The Chemistry of the Atmosphere and Oceans.* New York:Wiley Intersci.

Holland HD, Lazar B, McCaffrey M. 1986. *Nature* 320:27–33

Hoyle F. 1955. *Frontiers in Astronomy.* London:Heinemann

Hunten DM. 1973. *J. Atmos. Sci.* 30:1481–94

Hunten DM. 1993. *Science* 259:915–20

Ingersoll AP. 1969. *J. Atmos. Sci.* 26:1191–98

Jambon A, Zimmerman JL. 1990. *Earth Planet. Sci. Lett.* 101:323–31

Kahn R. 1985. *Icarus* 62:175–90

Kasting JF. 1988. *Icarus* 74:472–94

Kasting JF. 1989. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* (*Global and Planet. Change Sect.*) 75:83–95

Kasting JF. 1991a. In *Planetary Sciences, American and Soviet Research,* ed. TM Donahue, KK Trivers, DM Abramson, pp. 234–45 Washington, DC:Natl. Acad. Press

Kasting JF. 1991b. *Icarus* 94:1–13

Kasting JF, Ackerman TP. 1986. *Science* 234:1383–85

Kasting JF, Grinspoon DH. 1991. In *The Sun in Time,* ed. CP Sonett, MS Giampapa, MS Matthews, pp. 447–62 Tucson:Univ. Ariz. Press

Kasting JF, Pollack JB. 1983. *Icarus* 53:479–508

Kasting JF, Pollack JB, Ackerman TP. 1984. *Icarus* 57:335–55

Kasting JF, Toon OB, Pollack JB. 1988. *Sci. Am.* 258(2):48–53

Kasting JF, Whitmire DP, Reynolds RT. 1993. *Icarus* 101:108–28

Kearey P, Vine FJ. 1990. *Global Tectonics.* Oxford:Blackwell

Komabayashi M. 1967. *J. Meteorol. Soc. Jpn.* 45:137–39

Krumbein WE, Schellnhuber H-J. 1990. In *Facets of Modern Biogeochemistry,* ed. V Ittekkot, S Kempe, W Michaelis, A Spitzy, pp. 5–22. Berlin:Springer-Verlag

Kuhn WR, Atreya SK. 1979. *Icarus* 37:207–13

Larson Rl. 1991. *Geology* 19:963–66

Lewis J. 1982. *Icarus* 16:241–52

Lovelock JE. 1979. *Gaia.* Oxford:Oxford Univ. Press

Lovelock JE. 1989. *The Ages of Gaia.* Oxford:Oxford Univ. Press

Lovelock JE, Allaby M. 1984. *The Greening of Mars.* New York:Warner

Lovelock JE, Margulis L. 1974. *Tellus* 26:2–9

Lovelock JE, Whitfield M. 1982. *Nature* 296:561–63

Marshall HG, Walker JCG, Kuhn WR. 1988. *J. Geophys. Res.* 93:791–801

Matsui T, Abe Y. 1986a. *Nature* 319:303–5

Matsui T, Abe Y. 1986b. *Nature* 322:526–28

McElroy MB, Prather MJ, Rodriquez JM. 1982. *Science* 215:1614

McElroy, MB Kung T-Y, Yung YL. 1977. *J. Geophys. Res.* 82:4379–88

McElroy MB, Yung YL, Nier AO. 1976. *Science* 194:70–72

McGovern PJ, Schubert G. 1989. *Earth Planet. Sci. Lett.* 96:27–37

McKay CP. 1991. *Icarus* 91:93–100

McKay CP, Davis WL. 1991. *Icarus* 90:214–21

McKay CP, Haynes RH. 1990. *Sci. Am.* Dec:109–16

McKay CP, Toon OB, Kasting JF. 1991. *Nature* 352:489–96

Nakajima S, Hayashi Y-Y, Abe Y. 1992. *J. Atmos. Sci.* 49:2256–66

Newman MJ, Rood RT. 1977. *Science* 198:1035–37

Oberbeck VR, Marshall JR, Aggarwal H. 1993. *J. Geol.* 101:1–19

Ogawa M. 1993. *Icarus* 102:40–61

Opdyke BN, Wilkinson BH. 1988. *Paleoceanogr.* 3:685–703

Owen T, Cess RD, Ramanathan V. 1979. *Nature* 277:640–42

Parker TJ, Gorsline DS, Saunders RS, Pieri DC, Schneeberger DM. 1993. *J. Geophys. Res.* 98:11,061–78

Pepin RO. 1991. *Icarus* 92:2–79

Pollack JB. 1971. *Icarus* 14:295–306

Pollack JB. 1991. *Icarus* 91:173–98

Priem HNA. 1990. *Geol. Mijnbouw* 69:391–406

Prinn RG, Fegley B Jr. 1987. *Annu. Rev. Earth Planet. Sci.* 15:171–212

Rampino MR. 1992. *Eos, Trans. Am. Geophys. Union (Suppl)* 73:99

Rampino MR. 1994. *J. Geol.* In press

Rampino MR, Volk T. 1987. *Nature* 332:63–65

Rasool SI, DeBergh C. 1970. *Nature* 226:1037–39

Ronov AB, Yaroshevsky AA. 1976. *Geochem. Int.* 12:89–121

Sackman I-J, Boothroyd AI, Fowler WA. 1990. *Astrophys. J.* 360:727–36

Sagan C. 1960. *JPL Tech. Rep. No. 32-34*

Sagan C, Mullen G. 1972. *Science* 177:52–56

Savage MT. 1993. *The Millennial Project.* Denver:Empyrean

Schneider SH, Boston P, eds. 1991. *The Science of Gaia.* Cambridge:MIT Press

Schneider SH, Gal-Chen T. 1973. *J. Geophys. Res.* 78:6182–94

Schneider SH, Thompson SL. 1980. *Icarus* 41:456–69

Schwartzman D, Volk T. 1989. *Nature* 340:457–60

Sellers WD. 1969. *J. Appl. Meteorol.* 8:392–400

Tajika E, Matsui T. 1992. *Earth Planet. Sci. Lett.* 113:251–66

Tillman JE, Johnson NC, Guttorp P, Percival DB. 1993. *J. Geophys. Res.* 98:10,963–71

Volk T. 1987. *Am. J. Sci.* 287:763–79

Volk T. 1989a. *Geology* 17:107–10

Volk T. 1989b. *Nature* 337:637–40

Walker JCG. 1975. *J. Atmos. Sci.* 32:1248–56

Walker JCG. 1977. *Evolution of the Atmosphere.* New York:McMillan

Walker JCG, Hays PB, Kasting JF. 1981. *J. Geophys. Res.* 86:9776–82

Walter M. 1979. *Am. Sci.* 67:142

Watson AJ, Donahue TM, Walker JCG. 1981. *Icarus* 48:150–66

Wetherald RT, Manabe S. 1975. *J. Atmos. Sci.* 32:2044–59

Wetherill GW. 1986. In *Origin of the Moon,* ed. WK Hartmann, RJ Phillips, GJ Taylor, pp. 519–50. Houston:Lunar Planet. Inst.

Zahnle K, Kasting JF, Pollack JB. 1988. *Icarus* 74:62–97