

Xenological Constraints on the Impact Erosion of the Early Martian Atmosphere

KEVIN J. ZAHNLE

NASA Ames Research Center, Moffett Field, California

The roughly uniform hundredfold depletion of observable Martian atmophiles (nonradiogenic noble gases and reconstituted nitrogen) with respect to Earth implies that Mars lost its atmosphere by a relatively efficient, nonfractionating process. Impact erosion (expulsion of atmosphere by impacts) is an appealing candidate. Noble gases can be used to test this hypothesis. Xenon in particular can be used to impose three constraints on how Mars lost its atmosphere: its very low abundance compared to Earth, Venus, and likely meteoritic sources; its distinctive isotopic composition compared to likely meteoritic sources; and the relatively high absolute abundance of radiogenic ^{129}Xe , daughter of the extinct radionuclide ^{129}I (half-life 17 m.y.). A fourth useful constraint is imposed by radiogenic ^{40}Ar . If produced by impact erosion, the first two constraints become constraints on the composition, mass distribution, and orbital elements of the impactors. The third and fourth constraints imply that Mars lost its nonradiogenic noble gases early, perhaps before it was 100 Myr old. Impact erosion can be invoked to explain Mars by any of three stories: (1) Mars is unlikely. In a sort of planetary brinkmanship, impact erosion almost removed the entire atmosphere but was arrested just in time. (2) Martian noble gases are cometary and cometary Xe is as isotopically mass fractionated as Martian and terrestrial Xe. This is most easily accomplished if a relatively thick geochemically controlled CO_2 atmosphere protected trace atmophiles against escape. It is not known if comets actually have the desired composition. (3) Mars was indeed stripped of its early atmosphere but a small remnant was safely stored in the regolith, later released as a by-product of water mobilization.

INTRODUCTION

Mars is a small planet with a thin atmosphere. That Mars is small has been known for centuries, but proof of its thin atmosphere has been called "one of the great disappointments of the space age" [Anders and Owen, 1977, p. 453]. The apparent dearth of volatiles is not confined to one or two key elements but may well extend across the whole range of plausible atmospheric constituents. The most conservative estimates of the Martian volatile inventory resemble nothing so much as Earth depleted more or less evenly by a factor 30–100, a point duly noted by many workers in the aftermath of the Viking mission, and at the time interpreted in terms of a late accreting volatile-rich veneer shared unequally among the terrestrial planets [cf. Anders and Owen, 1977]. Severe depletion is surest for the noble gases, for which alternative reservoirs larger than the atmosphere are unlikely. It is less certain for less volatile volatiles, such as water or CO_2 , which can be hidden in surface and subsurface reservoirs as adsorbate, ice, hydrated minerals, or carbonate rocks. Still, the story of Mars would seem in large part the story of how it lost its atmosphere.

There is a widespread suspicion that Mars' fate is in some way attributable to its size. Three more or less complementary hypotheses have been suggested, all of which are plausible. A fourth hypothesis, that somehow Mars never came by volatiles in the first place, is hard to understand in the context of modern planetary accretion models [Wetherill, 1981], and is probably inconsistent with the SNC meteorites being from Mars [Dreibus and Wänke, 1985].

The most recently popular is that Mars, being small, cooled relatively quickly and long ago ceased to recycle

volatiles effectively. Weathering reactions of water, CO_2 , and rock would then have progressively and irreversibly consumed the atmosphere [Kahn, 1985]. Any excess water would have been permanently frozen deep in the regolith. Occasional volcanic episodes (or the slow brightening of the evolving Sun) would have released some of the hidden volatiles, the briefly liberated water carving the famous flood features, but overall the story is one of decay. The natural endpoint is Kahn's suggestion that the Martian atmosphere has equilibrated at the triple point of water, at which point weathering reactions sped by the occasional occurrence of liquid water suffice to offset a tiny CO_2 outgassing flux [Kahn, 1985]. This modern incarnation of Lowell's [1906] "dying planet" hypothesis applies to CO_2 and water in particular; it is more problematic for nitrogen, and it is very hard to see how it could apply to the noble gases.

The second story is that the thin Martian atmosphere is but a remnant of a once much thicker atmosphere, most of which long ago escaped to space [cf., Wallace, 1907]. Other things being equal, because it is smaller, escape is easier from Mars than from Earth or Venus. Several escape mechanisms have been suggested, including some that could be operative today. A possibly important example of the latter is the nonthermal escape of nitrogen [cf. Fox and Dalgarno, 1983]. Hydrodynamic escape and impact erosion of the atmosphere (a.k.a. atmospheric cratering) are two potentially much more effective escape mechanisms that should have been operative early. These matters will be discussed in more detail below.

The third hypothesis is that the atmosphere was never degassed or outgassed in the first place; i.e., the vast bulk of Martian atmophiles remain hidden in the mantle. Inefficient outgassing could be attributed to relatively low impact velocities during accretion (hence less impact degassing), relatively low energy of accretion (hence less in-

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ternal heating), and relatively high surface-volume ratio (hence rapid cooling and generally lower heat flow, less mantle turnover, and less volcanism). Inefficient outgassing of modern Mars is a reasonable hypothesis, and one that is fully consistent with the great antiquity of its few volcanos and strongly supported by the low absolute abundance of radiogenic ^{40}Ar compared to Earth [Scambos and Jakosky, 1990]. Relative abundances of radiogenic and non-radiogenic argon and xenon are shown in Figure 1. Inefficient exchange between internal and external reservoirs is also suggested by the remarkable differences between Martian internal rare gases (as represented by Chassigny [Ott, 1990]) and external rare gases, and has also been invoked to account for an inferred dessiccation of the Martian mantle [Carr and Wänke, 1992]. (Of course a dessiccated Martian interior does not imply that most Martian volatiles are hidden in the mantle, and it must be noted that there is no obvious way to derive Martian atmospheric noble gases from Chassigny.)

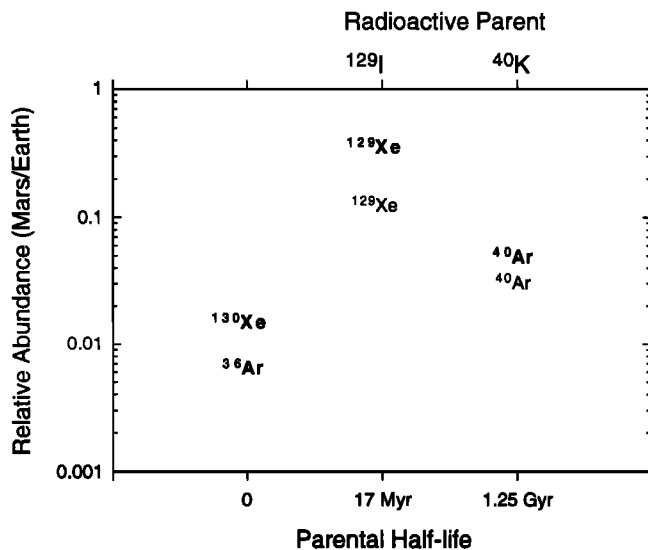


Fig. 1. Comparison of radiogenic and nonradiogenic Ar and Xe on Earth and Mars. The corrected (boldface) ratios allow for higher abundances of iodine and potassium in Mars than in Earth, as predicted by geochemical models [Dreibus and Wänke, 1989].

Nevertheless, inefficient outgassing of ancient Mars is problematic. For Earth, at least, early outgassing is a demonstrable fact, because the presence of excess ^{129}Xe in the mantle demands that Earth degassed before the parent ^{129}I (half-life 17 m.y.) became extinct [Ozima and Podosek, 1983]. Recent theoretical models of planetary accretion strongly imply that Mars, like Earth, should have accreted hot and when young should have convected itself with vigor [Schubert et al., 1992]. Since both heat flow and volatile transport in the mantle occur by convection, one might expect that Mars too would have outgassed early. On the other hand, if efficient outgassing of Earth were due less to convection than to bubble outgassing of a terrestrial magma ocean [Zhang and Zindler, 1989], then it is possible that Mars enjoyed a different fate. The relatively high abundance of ^{129}Xe in the Martian atmosphere can be taken as direct evidence in favor of relatively efficient early outgassing, although it could also simply indicate that the Martian atmosphere is a few tens of million of years older

than Earth's, thus recalling an age when ^{129}I was more abundant. On the whole it does not appear that the mantle is a likely hiding place for the missing volatiles on Mars.

The failure of the dying planet hypothesis to account for the noble gases tells us that it is at best incomplete. Something else is needed. The evidence preserved in the noble gases favors early escape.

The balance of this study examines impact erosion as the agent of escape. First I will address the several constraints placed on Martian atmospheric evolution by the noble gases, chiefly xenon. Then I will present a simple model describing the competition between impact erosion and impact supply of atmophiles. The model features four parameters: q , which describes the mass distribution of the impactors; y , their atmophile content; f , the fraction of impactors that strike slowly enough that impact erosion does not occur; and x , which represents the total mass of impacting material. The model is applied to two basic scenarios: a "one-volatile planet", in which all atmophiles behave like noble gases; and a "fixed-atmosphere planet", in which atmospheric CO_2 is controlled by a much larger geochemical buffer. Neither model can account for Mars in a straightforward manner. The study closes with a brief summary of some remaining possibilities.

THREE XENOLOGICAL CONSTRAINTS

Of the noble gases xenon imposes the most interesting constraints. Being the heaviest of the noble gases, xenon is the least susceptible to escape. Yet Martian nonradiogenic xenon is spectacularly rare compared to its abundance in carbonaceous chondrites, a volatile-rich class of meteorite often viewed as representative of the source of Martian geochemical volatiles, and in which xenon is, relative to Mars, much the most abundant noble gas. If the isotopic composition of xenon in carbonaceous chondrites were agreeable, the present Martian xenon inventory could be supplied by a veneer of carbonaceous chondritic material just 50 m thick. Although this is a large amount of material compared to the present Martian impact flux (of order 1 m over 3.5 b.y.), 50 m is much less than one might expect for the heavy bombardment (of order 1 km for Mars [Treiman et al., 1986]) and negligible compared to the mass of carbonaceous material plausibly delivered during the main phases of accretion. To illustrate the latter point: in order to account for the high abundance of geochemical volatiles seen in the SNC meteorites, Dreibus and Wänke [1987] propose that the material that accreted to form Mars was 40% carbonaceous chondritic. Mars retains about 1 part in 10^4 of the xenon that should have come with these carbonaceous chondrites. Or consider Mars today. Carbonaceous chondrites are estimated to comprise about 60% (by mass) of the fresh material presently reaching Earth [Shoemaker et al., 1990]; it is not unreasonable to assume that they comprised a fair fraction of the late heavy bombardment. In sum, there is far too little xenon on Mars to be explained by volatile-poor accretion alone.

Not only is Martian xenon scarce, it is isotopically unique (Figure 2). Its nonradiogenic isotopes, as inferred from trapped gases in the SNC meteorite EETA 79001 [Swindle et al., 1986], are strongly mass fractionated with respect to solar wind xenon or xenon in meteorites, including, in particular, the carbonaceous chondrites. In this respect Mar-

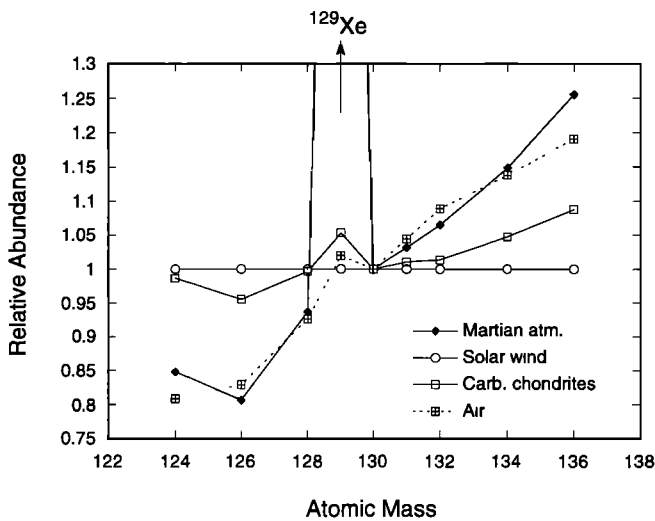


Fig. 2. Xenon isotopes on Mars compared to xenon in meteorites and the solar wind [after Zahnle, 1992]. Terrestrial xenon (air) is also mass fractionated to a similar degree.

tian atmospheric xenon resembles terrestrial xenon. Thus the second constraint is that the original Martian xenon must continue to look Martian through the ages, despite the probable late accretion of relatively large amounts of meteoritic xenon.

The origin of this xenon fractionation is not known. One interpretation is that the broad similarity between Earth and Mars implies that the two planets accreted xenon from a common source. We have elsewhere proposed a mechanism by which fractionated xenon would have been incorporated naturally into large ice-free planetesimals [Zahnle *et al.*, 1990b]. Another interpretation is that xenon fractionation occurred in situ on the two planets as a by-product of vigorous large-scale hydrodynamic hydrogen escape [Pepin, 1991]. This viewpoint regards the similar xenon fractionations as coincidental, instead emphasizing the isotopic differences between Xe on the two planets. In particular, Martian xenon looks like mass fractionated carbonaceous chondritic xenon, while Earth's looks like mass fractionated "U-Xe", a hypothetical primitive xenon relatively deficient in the heaviest xenon isotopes. Neither explanation is altogether satisfactory, for reasons which will not be discussed here [see Zahnle *et al.*, 1990b].

The third constraint, alluded to earlier, is the relatively high abundance of radiogenic ^{129}Xe on Mars. Figure 1 compares the relative abundances of nonradiogenic and radiogenic Xe and Ar on Earth and Mars. Figure 1 was prepared with SNC data, but need not have been — Viking data are quantitatively similar, if less precise. Radiogenic ^{40}Ar , daughter of ^{40}K (half-life 1.25 b.y.), is about 15 times less abundant on Mars than on Earth, as measured on a gram (argon) per gram (planet) basis. The apparent outgassing efficiency of ^{129}Xe , daughter of the extinct radionuclide ^{129}I , is relatively much higher, although still somewhat lower than Earth. About 7% of Earth's ^{129}Xe is radiogenic [Ozima and Podosek, 1983]. This corresponds to a radiogenic ^{129}Xe inventory of 6×10^{-15} g/g. By contrast Mars has an enormous ^{129}Xe excess of $\sim 260\%$ [Swindle *et al.*, 1986; Pepin, 1991], which corresponds to a radiogenic ^{129}Xe inventory of 2.2×10^{-15} g/g. In this one case the difference between Earth and Mars is less than a factor

3. When corrected for potassium and iodine abundances given by Dreibus and Wänke [1989], the Martian atmosphere appears a little less enriched in the radiogenic isotopes. Meanwhile, compared to Earth, Mars is seventyfold depleted in nonradiogenic xenon and 160-fold depleted in nonradiogenic argon.

The above observations imply that (1) either early outgassing efficiencies of Earth and Mars were not too different or that Mars's atmosphere is a few half lives of ^{129}I older than Earth's; (2) later outgassing of Mars was at best some 5 to 15 times less efficient than of Earth, most likely reflecting a failure of volcanism; (3) escape of the nonradiogenic noble gases from Mars (or at least xenon) occurred before the outgassing of much if not most of the ^{129}Xe ; and (4) escape must have been relatively unimportant by the time ^{129}Xe reached the atmosphere. The latter two points imply that Mars lost its nonradiogenic noble gases very early, probably before the extinction of the parent ^{129}I , and most likely within a hundred million years of the formation of the planet [cf. Musselwhite *et al.*, 1991].

THREE MORE XENOLOGICAL CONSTRAINTS

The constraints discussed above tell a tidy tale of atmospheres gained and atmospheres lost. But there are other radiogenic xenon isotopes that are not as easily understood as ^{129}Xe , and there is another kind of possibly Martian xenon that resembles the atmosphere not at all.

Because it is a single isotope, radiogenic ^{129}Xe is easily identified on a plot of xenon isotopes. Fission xenon is different. Fission xenon, which is formed by the spontaneous fission of ^{238}U (half-life 4.55 b.y.) or (extinct) ^{244}Pu (half-life 82 m.y.), is spread unevenly over the heavier stable isotopes ^{136}Xe , ^{134}Xe , ^{132}Xe , and ^{131}Xe [Ozima and Podosek, 1983]. It is also rare, because spontaneous fission is rare. These facts make fission xenon hard to separate from other sources of isotopic variation. At present, there is no evidence for fission xenon in Martian atmospheric gases [Swindle *et al.*, 1986].

Uranium fission xenon is too rare for its absence to serve as a useful diagnostic. Small amounts of fission xenon have been reported in some mantle samples [Allègre *et al.*, 1987]. The quantity and quality are consistent with ^{238}U being the parent. Even if all the ^{238}U fission xenon ever spawned in Earth were placed in the atmosphere, it would constitute less than 1% of atmospheric ^{136}Xe , which is not in practice discernable. Because Mars has only 1.5% as much nonradiogenic Xe as Earth, ^{238}U fission xenon would be obvious in the Martian atmosphere had it outgassed. The lack of detectable fission Xe on Mars is consistent with a low modern outgassing efficiency and the long lifetime of the parent.

The apparent absence of ^{244}Pu fission Xe on Mars is more interesting. As emphasized above, the Martian atmosphere is relatively rich in radiogenic daughters. It has been inferred that about 4% of Earth's ^{136}Xe is fissionogenic [Ozima and Igarashi, 1989; R. O. Pepin and D. Phinney, unpublished manuscript, 1978]. This inference is closely tied to the inferred composition of a hypothetical primordial terrestrial xenon called U-Xe. In these models Earth's Xe is obtained by mass fractionating a base of U-Xe by some 4% per AMU, and then adding some radiogenic ^{129}Xe and ^{244}Pu fission Xe. However, unlike ^{129}Xe , excess ^{244}Pu fission Xe has not been detected in mantle samples [Ozima

and Igarashi, 1989]. This poses a puzzle. The existence of excess ^{129}Xe demands that ^{129}I was alive when the air separated from the mantle. Because ^{244}Pu enjoys a longer half-life than ^{129}I and plutonium is less volatile than iodine, it must be expected that fissionogenic Xe should be more excessive in mantle samples than ^{129}Xe . Yet such is not the case. It is possible that Earth's plutonium was somehow sequestered at some early time. Another possibility is to omit ^{244}Pu fission Xe as a fundamental component of Earth's atmosphere. It is far from obvious what its absence means for Mars.

The most puzzling Martian rare gases are those contained in the Chassigny meteorite [Ott, 1988]. Chassigny, the "C" in "SNC", is thought by some [e.g. Pepin, 1991] to sample internal Martian gases. If so, it reveals a relationship between mantle and atmosphere utterly unlike Earth. Earth's mantle is enriched in radiogenic isotopes relative to air, consistent with the mantle being the source of the radiogenic isotopes. Otherwise mantle Ar, Kr, and Xe are isotopically like air [Ozima and Igarashi, 1989]. By contrast, Chassigny lacks radiogenic rare gases. Therefore Chassigny does not sample the source of radiogenic noble gases in the Martian atmosphere. Another contrast is that Chassigny Xe is isotopically similar to that in carbonaceous chondrites or the solar wind, not the Martian atmosphere. A third oddity of Chassigny is that its Kr/Xe and Ar/Xe ratios are very low, again more like carbonaceous meteorites than the Martian atmosphere.

If Chassigny noble gases do indeed represent a part of the Martian mantle one must conclude that outgassing of this part of the mantle has had little impact on the composition of the Martian atmosphere. The absolute abundance of Xe is high enough in Chassigny (^{130}Xe at $\sim 4 \times 10^{-14}$ g/g [Ott, 1988]) and its isotopic composition so discordant with the atmosphere that outgassing from no more than ~ 800 m of Chassigny-like volcanism can be comfortably hidden in the Martian atmosphere. Over the past 3.8 b.y. this corresponds to magmas mounting at less than about $0.03 \text{ km}^3 \text{ yr}^{-1}$. By contrast, Greeley and Schneid [1991] estimate that over that same period the average amount of "extrusive and intrusive magma generated on Mars" has been $0.17 \text{ km}^3 \text{ yr}^{-1}$. Of course the xenon still trapped within Chassigny is xenon that was not outgassed, a caveat that clouds interpretation of 800 m of Chassigny-like volcanics as an upper limit.

MEANS OF ESCAPE

Hydrodynamic escape refers to thermal escape when escape fluxes are high (classical Jeans escape applies to thermal escape at low fluxes). In hydrodynamic escape the upper atmosphere simply flows into space, much like the solar wind. For this to occur from a terrestrial planet a hydrogen-rich atmosphere is probably required; such is not unreasonable during accretion. Large-scale hydrodynamic hydrogen escape could have stripped Mars of much of its initial endowment of volatiles [Dreibus and Wänke, 1987; Pepin, 1991]. Because hydrodynamic escape is inherently mass fractionating, it may have contributed to the observed high D/H ratio [Owen *et al.*, 1989], and it has also been invoked to explain Mars' high $^{36}\text{Ar}/^{36}\text{Ar}$ ratio [Zahnle *et al.*, 1990a] and the fractionation pattern of the Martian xenon isotopes [Pepin, 1991]. Mass fractionation by hydrodynamic escape could also account for the observed modest elemen-

tal fractionation of the Martian noble gases with respect to Earth [Hunten *et al.*, 1987]; however, hydrodynamic escape alone cannot easily explain both the elemental and the isotopic fractionations, since the relatively strong isotopic fractionations are incompatible with the relatively weak elemental fractionation. Although in principle hydrodynamic escape could have depleted Mars's atmosphere thirtyfold while leaving the composition of the remnant atmosphere nearly unaltered, very high escape fluxes are required.

The other effective theoretical tool for prying volatiles off a planet is impact erosion, which refers to the expulsion of atmosphere by impacts [Melosh and Vickery, 1989]. Because it is likely to have been efficient, nearly non-fractionating, and more potent for Mars than for Earth or Venus, impact erosion appears to be a better candidate than hydrodynamic escape for converting an Earth-like Mars to a Mars-like Mars. Because volatiles are lost in proportion to the impact flux, impact erosion should have been most effective within the first hundred million years. I will emphasize impact erosion in this essay, and in particular its effect on the noble gases, since they are the most volatile of the elements, and so are both the most likely to have been affected by impact erosion and the easiest to address quantitatively. However, both impact erosion and hydrodynamic escape are plausible, and modern Mars may have lost a great deal to each.

Atmospheric Cratering

Melosh and Vickery [1989] proposed that atmospheric gases are swept to space by the momentum of a hydrodynamically expanding cloud of superheated vapor (and particles that condense from it) produced by impact. Escape occurs if the cloud attains escape velocity and if it has enough extra momentum to carry the intervening atmosphere with it. Melosh and Vickery's tangent-plane approximation is to assume that when atmospheric cratering occurs, all of the atmosphere above the horizon escapes; i.e., that part of the atmosphere lying above a plane tangent to the surface of the planet escapes. Escape occurs if (1) the expansion velocity of the rock vapor exceeds the escape velocity from the planet, and (2) the mass of rock vapor exceeds the mass of atmosphere above the tangent-plane.

Melosh and Vickery suggested a threshold impact velocity v_c for atmospheric cratering of

$$v_c = 2\sqrt{v_{\text{esc}}^2 + 2L_v}, \quad (1)$$

where v_{esc} is the escape velocity and L_v is the latent heat of vaporization (for which Melosh and Vickery use 1.3×10^{11} ergs/g for rock and 3×10^{10} ergs/g for ice). Equation (1) assumes that the impactor and the target are composed of the same material and that all the internal energy deposited by the shock becomes thermal energy that powers escape. For unlike materials the fraction of the impactor's kinetic energy deposited as internal energy in the target is

$$\frac{u_t}{u_i} \left(1 + \frac{u_t}{u_i}\right)^{-2} \quad (2)$$

and the fraction deposited as internal energy in the impactor is

$$\frac{u_t^2}{u_i^2} \left(1 + \frac{u_t}{u_i}\right)^{-2}, \quad (3)$$

where the subscripts “t” and “i” denote the target and the impactor, respectively [Melosh, 1989, p. 57]. The velocities u_i and u_t are the postshock particle velocities measured in the rest frame of the relevant unshocked material; thus u_i is measured in the rest frame of the impactor. In the strong shock limit

$$\rho_t u_i^2 \approx \rho_i u_t^2 \quad (4)$$

holds approximately. Modified expressions for v_c for unlike materials that result are

$$v_{ci}^2 \equiv \frac{(1 + \sqrt{\rho_i/\rho_t})^2}{h} (v_{esc}^2 + 2L_{vi}) \quad (5)$$

for the escape of the impactor and

$$v_{ct}^2 \equiv \frac{(1 + \sqrt{\rho_t/\rho_i}) (1 + \sqrt{\rho_i/\rho_t})}{h} (v_{esc}^2 + 2L_{vt}) \quad (6)$$

for the escape of a mass of target comparable to the mass of the impactor. Equations (5) and (6) also assume that only a fraction $0 < h < 1$ of the internal energy deposited by the shock is ultimately channelled into escape. Schmidt and Hausen [1987] crater scaling for a porous medium (dry sand) is equivalent to taking $h = 0.8$; for a non-porous medium (wet sand) $h = 0.33$ [Zahnle, 1990]. Melosh and Vickery’s criterion is recovered by setting $h = 1$.

Table 1 lists some representative threshold impact velocities for escape from Mars. These idealize the impact of comets and asteroids on icy and rocky surfaces by setting $\rho = 3 \text{ g/cm}^3$ for rock and $\rho = 1 \text{ g/cm}^3$ for ice. Note that comets are more erosive than asteroids and volatiles borne by comets are harder to retain than volatiles borne by asteroids, even if impact velocities are equal. Because (2) is derived for plane-parallel impact it underestimates the mass of target material that is shocked during the early stages of a three-dimensional impact, and therefore (2) overestimates the average energy deposited per unit mass. Consequently (6) probably underestimates the velocity threshold for the escape of target material.

TABLE 1. Threshold Impact Erosion Velocities for Mars

h	1	0.8	0.5	0.33
	<i>Rock on Rock</i>			
v_c	14.4	16.1	20.4	25.0
	<i>Ice on Ice</i>			
v_c	11.3	12.7	16.0	19.6
	<i>Rock on Ice</i>			
v_{ci}	19.7	22.0	27.9	34.1
v_{ct}	11.7	13.1	16.6	20.4
	<i>Ice on Rock</i>			
v_{ci}	8.9	10.0	12.6	15.5
v_{ct}	15.0	16.7	21.2	25.9

In km/s.

If the mass of the impactor exceeds the mass of atmosphere above the horizon, Melosh and Vickery assumed that there is enough momentum in the rock vapor to blow off all the atmosphere encountered. The fraction of an exponential atmosphere above a plane tangent to the surface is $\xi = H/2R$, where H is the atmospheric scale height and R is the planet’s radius. Escape occurs if the impactor mass m exceeds the threshold m_c , defined by

$$m \geq m_c = \frac{H}{2R} MY, \quad (7)$$

where Y represents the mass of the atmosphere normalized to the mass of the planet (grams of atmosphere per gram of planet). For the present Martian atmosphere $m_c \approx 4 \times 10^{16} \text{ g}$. This impactor mass corresponds to a 20–40 km diameter crater, depending on the impact velocity and the particular crater scaling relation used. Equation (7) is, for the atmosphere, potentially catastrophic. It states that a thinner atmosphere can be eroded by smaller impactors. Since smaller impacts are more numerous, a thinner atmosphere erodes more quickly. Thus once an atmosphere begins to erode, impact erosion proceeds until the planet is wholly stripped.

In reality, there will be a range of impacts for which a fraction of the mass above the horizon is expelled. Thus as a prescription for the effect of an individual impact the tangent-plane approximation can be misleading. But when integrated over a realistic spectrum of impactor masses the tangent-plane prescription appears to give a fair approximation to more sophisticated prescriptions that compare the momentum of the plume to the mass swept up along different slant paths through the atmosphere [cf., Vickery and Melosh, 1990]. The approximation breaks down when high velocity ($v_i > v_c$) impacts are rare (less than 10%). When most impacts are slow impact erosion is dominated by more massive impactors ($m \gg m_c$), from which only the relatively small fraction of high velocity vapor is needed to remove the intervening atmosphere. Consequently the effective impact erosion rate is unlikely to become vanishingly small, although the fraction of delivered volatiles lost on impact can become negligible. Another question is whether the escape velocity is the appropriate threshold velocity for escape; i.e., how high and how thin must the plume get to ensure that its entrained air escapes?

Impact Velocities on Mars

The ability of Mars to retain an atmosphere against impact erosion is determined by the mass distribution, velocity distribution, and volatile content of the impactors. Modern Martian impactors can be counted and assessed but little can be said with confidence about the late heavy bombardment, and still less about earlier objects that might more properly be regarded as the last chapter of planetary accretion. For want of better information it will be assumed here that the composition, mass distributions, and orbital distributions of ancient impactors were like those of the three important kinds of impactors today: Mars-crossing asteroids, short-period (Jupiter family) comets, and long-period (Oort Cloud) comets. This is probably a pretty good assumption for the orbital distributions, although the relative importance of the three types of orbit may have varied over time. The mass distribution of asteroids is probably collisional; if so, the modern distribution should also describe ancient asteroids. The assumption of similar composition is more questionable. Presently some 20% of the Mars-crossing asteroids are carbonaceous [C. Chyba, pers. comm.]. What it was in the past is unknown. There are whole regions of the asteroid belt from which we have no meteoritic samples, and ancient asteroids may have fallen from relatively unstable regions of the asteroid belt that were long ago depopulated. Meanwhile, the mass

distribution of comets is almost as badly known as their composition.

Figure 3 shows cumulative impact velocity distributions for asteroids and comets striking Mars. It is not meant to be definitive. Impact velocities and impact probabilities for asteroids and short-period comets are taken directly from tabulations by *Steel* [1985] and *Olsson-Steel* [1987], respectively. *Steel* lists both maximum and minimum encounter velocities; I have used both, weighting them equally. Impact velocities and probabilities for Oort Cloud comets are generated from synthetic orbits using *Öpik's* method [*Shoemaker and Wolfe*, 1982], assuming an isotropic source (uniform in $\sin i$, where i is inclination) and a uniform distribution of perihelia. Aphelia are placed at 20000 AU.

What is most noteworthy about the impact velocity distribution of asteroids is how slow they are. Most asteroid impacts on Mars are too slow to vaporize much rock or cause impact erosion of an atmosphere. This is largely because most Mars-crossing asteroids come from the innermost part of the asteroid belt, with perihelia just inside Mars's aphelion. As a consequence encounter velocities are generally quite low. Comparison of Figure 3 with impact velocities listed in Table 1 indicates that some 85-95% of incoming asteroids strike too slowly to expell much atmosphere. It is otherwise for SP comets, of which only some 10-40% would be expected to be retained (the precise value is rather sensitive to the low velocity tail of the distribution and to the parameter h). Oort Cloud comets are probably hopeless as a volatile source, with retention of no more than 1% of the impacting objects.

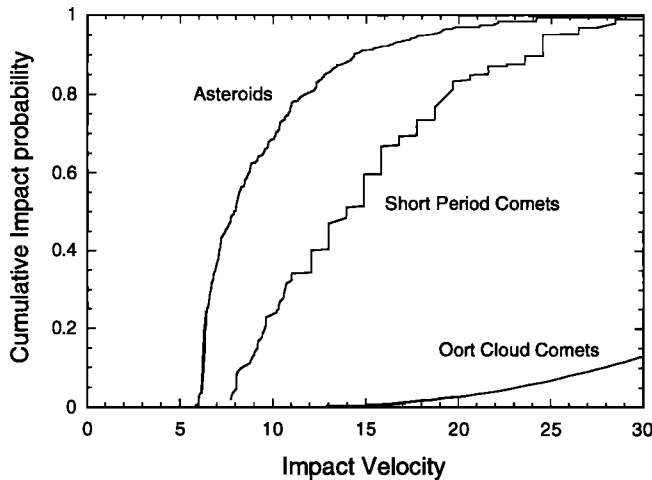


Fig. 3. Cumulative impact probability as a function of impact velocity for three modern solar system populations. The curves labeled "asteroids" and "short-period comets" use impact probabilities for the known populations of Mars-crossing objects. The other "Oort Cloud comets" are generated from synthetic distributions of orbital elements.

ATMOSPHERIC EVOLUTION CONTROLLED BY IMPACTS

The competition between accretion and impact erosion of atmospheres was addressed by *Zahnle et al.* [1992] in the context of the accumulation of Titan's atmosphere. Following *Zahnle et al.* [1992], let y_j denote the abundance of an atmospheric species j in grams per gram of impacting material, and let Y_j denote the atmospheric inventory of j in grams per gram of planet. An impactor mass flux \dot{M} is in-

cident on a planet of mass M_p . With outgassing neglected, the atmospheric inventory of j evolves according to

$$\frac{d}{dt}(M_p Y_j) = y_j \iint \dot{n}(m, v) \zeta(m, v) m \, dm \, dv - M_p Y_j \iint \dot{n}(m, v) \xi(m, v) \, dm \, dv. \quad (8)$$

The integrals are taken over all impactors. The first term on the right-hand side of (8) describes impact supply of volatiles. The function $\zeta(m, v)$ is the fraction of a particular impactor of mass m and impact velocity v retained by the planet, and $\xi(m, v)$ is the fraction of the planet's volatile inventory expelled by a given impact. The second term represents atmospheric cratering. For an oblique atmosphere, like a noble gas, ξ is the fraction of the atmosphere expelled in an impact, while for a condensed volatile, like an ocean, ξ would refer to the fraction of the ocean expelled. Since condensed volatiles are more likely to be retained, atmospheric cratering discriminates between strict atmophiles like the noble gases and probably nitrogen on the one hand, and water and possibly carbon dioxide on the other.

For a thin veneer ($\delta M \ll M_p$) it is acceptable to regard M_p as approximately constant. It is convenient to define an average retention factor χ and an average erosion efficiency η such that (8) becomes

$$\dot{Y}_j = \frac{\dot{M}}{M_p} (\chi y_j - (1 + \eta) Y_j) \approx \frac{\dot{M}}{M_p} (\chi y_j - \eta Y_j). \quad (9)$$

In (9) the factor $0 \leq \chi \leq 1$ is the net fraction of all the impactors' volatiles released into the atmosphere and retained by the planet. The cumulative effects of atmospheric cratering are integrated into a single nondimensional factor η , defined as the ratio of the planet's accretion time scale to the atmosphere's escape time scale.

The (differential) mass spectrum $\dot{n}(m, v)$ is assumed to obey a power law

$$\dot{n}(m, v) \, dm \propto m^{-q} \, dm. \quad (10)$$

Popular values of q range from about 1.5 to 2. These include 1.83 for a (theoretical) distribution dominated by fragmenting collisions [*Dohnanyi*, 1972]; 1.54 from inverting the lunar cratering record [*Chyba*, 1991]; ~ 2 for large asteroids [*Hughes*, 1982; *Donnison and Sugden*, 1984]; ~ 1.7 for long-period comets [*Hughes*, 1988; *Donnison*, 1986]; and 1.45 for short-period comets [*Donnison*, 1986], although the cometary data, at least, may admit other interpretations [*Weissman*, 1991]. None of these values should be taken too seriously. I will treat q as a free parameter. The velocity spectra were discussed above in connection with Figure 3.

The distribution in (10) is normalized by integrating over all m to get the mass accretion rate \dot{M} . For $q < 2$,

$$\dot{n}(m, v) \, dm = \frac{(2-q)\dot{M}}{m_1^{2-q}} m^{-q} \, dm, \quad (11)$$

where the largest impactor in the distribution is denoted m_1 . Since our primary interest is a late veneer, it is useful to define the mass of veneer incident after time t as

$$\delta M(t) \equiv \int_t^{\text{present}} \dot{M}(t) \, dt. \quad (12)$$

Note that as defined here, the veneer δM decreases with time. It should also be noted that with impact erosion, the actual accreted mass is less than the incident mass δM . Indeed, it is entirely possible for a planet or moon to shrink under high velocity bombardment. The mass of the largest impactor m_1 in the veneer is statistically related to the total mass δM of the veneer [Wetherill, 1975]

$$m_1 \approx \left(\frac{4-2q}{3-q} \right) \delta M. \quad (13)$$

In working with (9) it is convenient to define a nondimensional veneer mass

$$x \equiv \frac{\delta M}{M_p}. \quad (14)$$

Consistent with the tangent plane approximation it is assumed that all objects with impact velocities greater than v_c and masses greater than m_c are erosive, with expulsion both of the atmosphere above the tangent plane and of the impactor itself; otherwise no erosion occurs and the impactor's volatiles are contributed to the atmosphere. Then, with y_j assumed constant, the average source efficiency χ is

$$\chi = f + (1-f) \left(\frac{3-q}{4-2q} \xi Y \right)^{2-q} x^{q-2}, \quad (15)$$

where the f is the fraction of impactors with impact velocity less than v_c . The first term is the fraction of all impactors that hit slowly enough that their volatiles are retained. The second term refers to impactors that are too small for their ejecta to escape: either they are stopped in the atmosphere, or their ejecta are smothered by the atmosphere. To the same approximation the atmospheric cratering efficiency η is

$$\eta = (1-f) \frac{2-q}{q-1} \left(\frac{3-q}{4-2q} \xi \right)^{2-q} Y^{1-q} x^{q-2}. \quad (16)$$

The above expressions assume that $(m_1/m_c)^{q-1} \gg 1$.

When written out, (9) for the inventory of the atmosphere j based on the tangent-plane model for atmospheric cratering becomes

$$\begin{aligned} \frac{dY_j}{dx} = & -fy_j - (1-f) \left(\frac{3-q}{4-2q} \frac{H}{2R} \right)^{2-q} Y^{2-q} y_j x^{q-2} \\ & + (1-f) \frac{2-q}{q-1} \left(\frac{3-q}{4-2q} \frac{H}{2R} \right)^{2-q} Y^{1-q} Y_j x^{q-2}. \end{aligned} \quad (17)$$

In order, the three terms on the right-hand side represent (1) volatiles contributed by slow impactors, (2) volatiles contributed by small, fast impactors, and (3) impact erosion by large, fast impactors. An additional relation $\sum Y_j = Y$ closes the system of equations represented by (17). Analytical solutions can be obtained given a simple enough prescription for either the evolving mass of the background atmosphere $Y(x)$ or for the mixing ratio Y_j/Y , or if (17) is further simplified [Zahnle et al., 1992]. Examples of the former are to assume constancy of Y or of Y_j/Y . An example of the latter is to assume that the impactors carry no volatiles, i.e., $y_j = 0$, which is useful for estimating the maximum atmosphere that can be completely expelled by a given mass veneer and also for direct comparison with the work of Melosh and Vickery [1989].

I will consider two idealized kinds of impact erosion. In

the first, all atmospheric constituents, including CO_2 , will be treated as if they were noble gases. Thus the planet is regarded as a kind of inert platform. In the second the primary atmospheric constituent (CO_2) is held constant, as might be the case if its abundance in the atmosphere were controlled geochemically. Implicit in this scenario is a large reservoir outside (beneath) the atmosphere which can be tapped at need.

The One-Volatile Planet

As a first approximation, consider the one-volatile planet. In this approximation CO_2 is treated as an inert gas that accumulates around an impermeable, unreactive planet. More generally, a one volatile planet is equivalent to assuming constant Y_j/Y , which should hold for all volatiles if the atmosphere is their primary reservoir and if the incoming volatiles have the same composition as the atmosphere, since to first approximation atmospheric cratering does not discriminate between atmospheric constituents in a well-mixed atmosphere: either a molecule is above the horizon or it isn't. However, it would not hold if the major atmospheric constituent had its primary reservoir outside the atmosphere and exchange was rapid. An idealized example of the latter case will be discussed in the next section. For the one-volatile planet, (17) reduces to

$$\frac{dY}{dx} = AY^{2-q} x^{q-2} - fy \quad (18)$$

for Y , the atmosphere as a whole, where the parameter

$$A \equiv (1-f) \left(\frac{2-q}{q-1} - y \right) \left(\frac{H}{2R} \frac{3-q}{4-2q} \right)^{2-q}. \quad (19)$$

Consider first a highly erosive case. With no incoming volatiles ($y = 0$) or if all large impacts are erosive ($f = 0$), the integral of (18) is

$$Y^{q-1}(t) = Y^{q-1}(t_0) + A(x^{q-1} - x_0^{q-1}). \quad (20)$$

The initial state is at $t = t_0$ and $x_0 \equiv x(t_0)$. Since $x(t) \rightarrow 0$ as t approaches the present, retention of a finite atmosphere places an upper bound on the mass of eroding veneer that can be accreted. With Y_f denoting the final atmosphere left after impact erosion is complete, the original atmosphere before erosion by a veneer of mass $x_0 M_p$ is

$$Y_0 = (Y_f^{q-1} + Ax_0^{q-1})^{1/(q-1)}. \quad (21)$$

The minimum veneer mass x_0 needed to strip a planet of an atmosphere Y_0 is

$$x_0 = A^{-1/(q-1)} Y_0. \quad (22)$$

Figure 4 shows the biggest atmosphere Y_0 that can be completely removed from Mars by volatile-poor ($y = 0$) "cometary" ($q = 1.7$, $f = 0.1$) and "asteroidal" ($q = 1.8$ and $q = 1.5$, $f = 0.85$) veneers of mass x_0 . It should be recalled that x_0 refers to the mass of impactors incident on the planet. The difference between a Y_0 that gives $Y_f = 0$ and a Y_0 that gives $Y_f = 3.6 \times 10^{-8}$ (modern Mars) is a measure of the likelihood of a thin atmosphere developing by impact erosion alone.

According to Figure 4, the present Martian atmosphere would be removed by a rocky veneer less than 10 m thick. That a mere 10 m of late accreting material could serve to

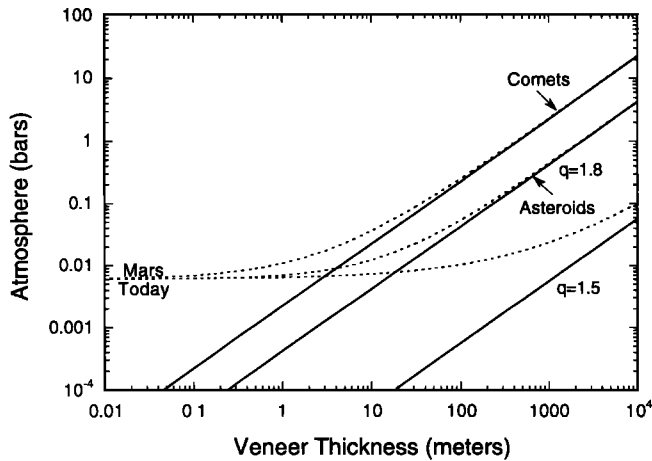


Fig. 4. Maximum impact erosion for Mars. The curves (solid and dashed) labeled “comets” use $q = 1.7$ and $f = 0$; the curves labeled “asteroids” use $f = 0.7$ and two values of q : 1.8 and 1.5. The solid curves show the largest atmosphere that can be removed by a late veneer of the given thickness ($\rho = 2.5 \text{ g/cm}^3$). The dashed curves show veneers that produce a modern Martian atmosphere for the same assumptions.

separate Mars from its present atmosphere suggests that the mass of the present Martian atmosphere is not indicative of its mass in the deep past. For comparison, the modern impacting flux is of the order of a meter over 4 b.y. Ancient impacts were of course more numerous. Based on the amount of iridium in the lunar crust, the Moon must have accreted at least 700 m of chondritic material since its crust stabilized ~ 4.4 b.y. [Sleep *et al.*, 1989]. The same argument applied to the SNC parent body (using rhenium) implies that Mars accreted a chondritic veneer no thicker than 1 km [Treiman *et al.*, 1986]. Thus there were probably enough late impacts to remove at least a 1-bar atmosphere from Mars. Impact erosion would have been an important process in the atmosphere under which the present Martian surface formed. However, this conclusion is sensitive to q for $q < 1.7$. Lower values of q describe impacting populations that are increasingly deficient in the relatively small impactors that are the bane of atmospheres. If the impactors of the late bombardment fell from a mass spectrum with $q \approx 1.5$, as has been suggested by Melosh and Vickery [1989] and Chyba [1991], impact erosion would have been relatively ineffective during the period of Mars’ geological record. This possibility is also depicted on Figure 5. But it should be remembered that the late veneer responsible for impact erosion and the late heavy bombardment responsible for the modern surface are not the same thing. Impact supply and impact erosion of volatiles occurs throughout the growth of the planet. It is wrong to imagine that impact erosion would have been restricted to the late heavy bombardment.

Solutions to (17) for the one-volatile planet in the more general case where $y \neq 0$ are discussed by Zahnle *et al.* [1992] in the context of Titan, Ganymede, and Callisto. Qualitatively, if impacts are important there are two types of solutions, depending on whether impact erosion or impact supply of volatiles gets the upper hand. In the first category a preexisting atmosphere is eventually eroded to nothing. These can be regarded as “erosive” histories. The atmosphere tends to retain its original composition as it

shrinks. In the second a preexisting atmosphere is supplemented by and eventually supplanted by the volatiles carried by the impactors. These may be regarded as “accumulative” histories. In these the atmosphere ultimately takes on the composition of the impactors.

The division between accumulative and erosive regimes for the one-volatile planet is the surface in (q, y, f) space determined by the relation [Zahnle *et al.*, 1992]

$$(2 - q) A = \left(\frac{2 - q}{q - 1} f y \right)^{q-1} \quad (23)$$

Figure 5a shows solutions to (23) for Mars for three values of f corresponding approximately to asteroids, SP comets, and Oort Cloud comets. An atmosphere would accumulate over an initially airless Mars if the average impactor atmosphere content plots above the relevant curve. If most of the carbon in carbonaceous asteroids were to reach the atmosphere, it is probable that for asteroids y would exceed 0.01, and so asteroids are likelier than not to be net sources of Martian atmosphere. Short period comets are also likelier to be sources than not, since their net CO_2 content (or the equivalent after shocking) is likely to exceed 10%. By contrast Oort Cloud comets are almost certainly a net sink of Martian atmospheres. It should also be noted that the likelihood of the late veneer being a net sink of atmospheres on Mars depends strongly on the fate of carbon carried by large, slow-moving impactors, since these by mass contribute the bulk of the veneer.

Figure 5b views solutions to (23) from another angle. Here solutions are plotted for four values of the impactor atmosphere content y . Mars accumulates an atmosphere if f for the impactors lies above the curve. Approximate locations of asteroids, short-period comets, and long-period comets on the q, f plane are indicated. The location of the late veneer on the q, f plane is not known.

Xenon and the One-Volatile Planet

My present purpose is to use Martian Xe to limit the role of carbonaceous chondrites. Because the isotopic compositions of impactor (y_i) and atmosphere (Y_i) differ, (17) must be solved numerically. The initial Martian atmosphere and the thickness of the late veneer impinging upon it must both be regarded as free parameters. Neither is well constrained. As a working hypothesis it will be assumed that the initial atmosphere was comparable to the atmospheres of Venus and Earth today. For carbon, nitrogen, and Xe these planets are the same to within roughly a factor 3 (Table 2).

Asteroids. As a first example (Figure 6a-c), assume that the impacting asteroids are described by $f = 0.7$ and $q = 1.8$, with average ^{130}Xe and carbon (dioxide) contents of $y_{130} = 1.4 \times 10^{-12} \text{ g/g}$ and $y_{44} = 0.01$, respectively. This composition assumes that the impactors are 20% carbonaceous chondritic. A ^{130}Xe content of $7 \times 10^{-12} \text{ g/g}$ former is the average for carbonaceous chondrites [Pepin, 1991]. The CO_2 composition implicitly assumes that much of the carbon carried by the impactors eventually reaches the atmosphere as CO_2 . This may be optimistic. The particular value $f = 0.7$ was chosen because with $x_0 = 0.01$ it erodes into a Mars-like atmosphere.

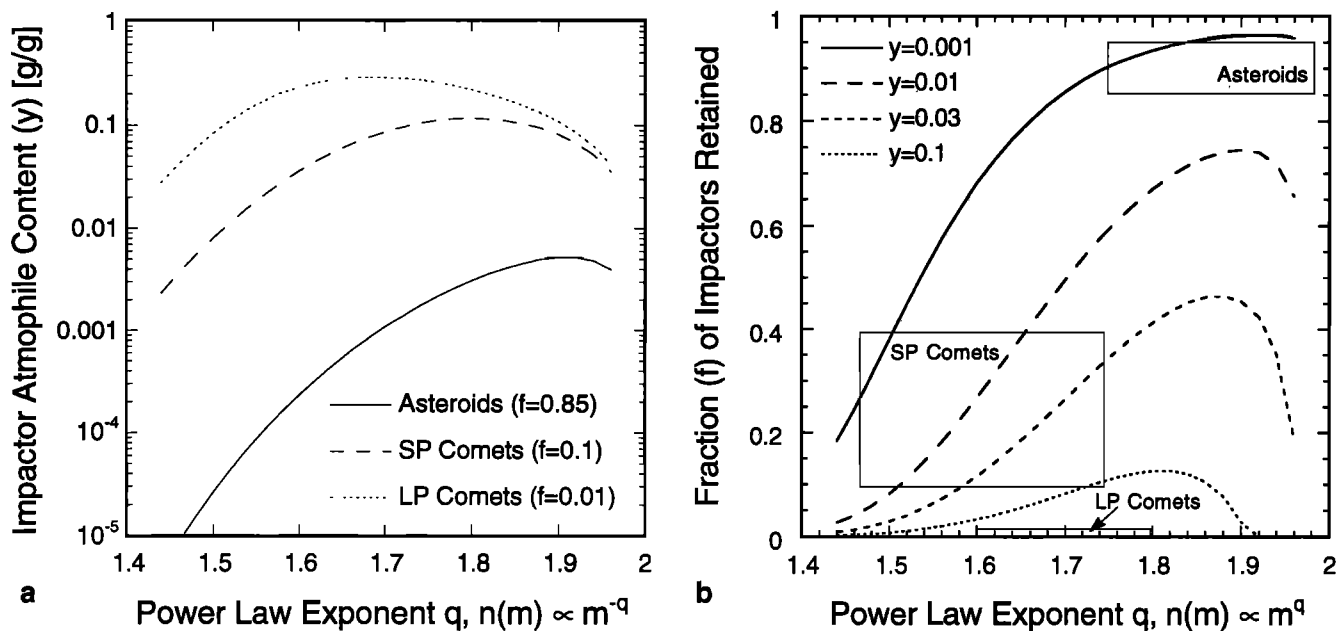


Fig. 5. Why is there air? (a) Mars will tend to accumulate an atmosphere if the average impactor atmosphere content is high enough to plot above the relevant curve. If the impactor atmosphere content plots below the curve, the impactors are fundamentally erosive. (b) The same question answered from another angle. Curves are marked by atmosphere content y . Mars accumulates an atmosphere if the impactors are slow enough for f to lie above the indicated curve. Approximate locations of asteroids, short-period comets, and long-period comets on the q, f plane are indicated.

TABLE 2. Superficial Volatile Inventories

	Venus	Earth	Mars
N_2	$2.2 \pm 0.5 \times 10^{-6}$	$9.1 \pm 2.4 \times 10^{-7}$	$7.3 \pm 1.9 \times 10^{-10}$
CO_2	$9.6 \pm 0.2 \times 10^{-5}$	$3.7 \pm 2.5 \times 10^{-5}$	$4.1 \pm 0.7 \times 10^{-8}$
^{130}Xe	$9.0 \pm 7.0 \times 10^{-14}$	$1.4 \pm 0.1 \times 10^{-14}$	$2.1 \pm 0.4 \times 10^{-16}$

Martian CO_2 and N_2 are lower limits. In g/g. After Pepin [1991].

Figure 6a shows evolutionary trajectories for a CO_2 atmosphere under asteroidal bombardment. Time goes to the right as $x \rightarrow 0$. Twelve cases are shown, six for a ~ 15 km veneer ($x_0 = 0.01$, solid curves) and six for a ~ 1.5 km veneer ($x_0 = 0.001$, dotted curves). The six cases correspond to different initial atmospheres, ranging from 1 mbar to 100 bars. The initially thicker atmospheres are not significantly eroded: the incident veneers are too thin. The initially thinner atmospheres shrink more noticeably, although for these particular parameters none vanish. For a given x_0 the different evolutionary paths converge to a common final atmosphere that represents an equilibration with the impactors.

Evolutionary tracks for ^{130}Xe are shown in Figure 6b for the same 12 cases. The initial ^{130}Xe mixing ratio is the same as modern Mars for all 12 cases (models starting with terrestrial Xe/CO_2 differ little). Most of the ^{130}Xe is lost, leaving a final abundance that is to order of magnitude comparable to that of Mars. With some fine tuning of the various parameters there is no doubt that a pretty close fit to modern Mars could be achieved.

However, the isotopic composition of the xenon poses a problem. Those atmospheres that equilibrate with the asteroids are also those that adopt the composition of the asteroids. This is shown in Figure 6c, a generalization of Figure 6b to a wide range of veneer thicknesses. Figure 6c plots

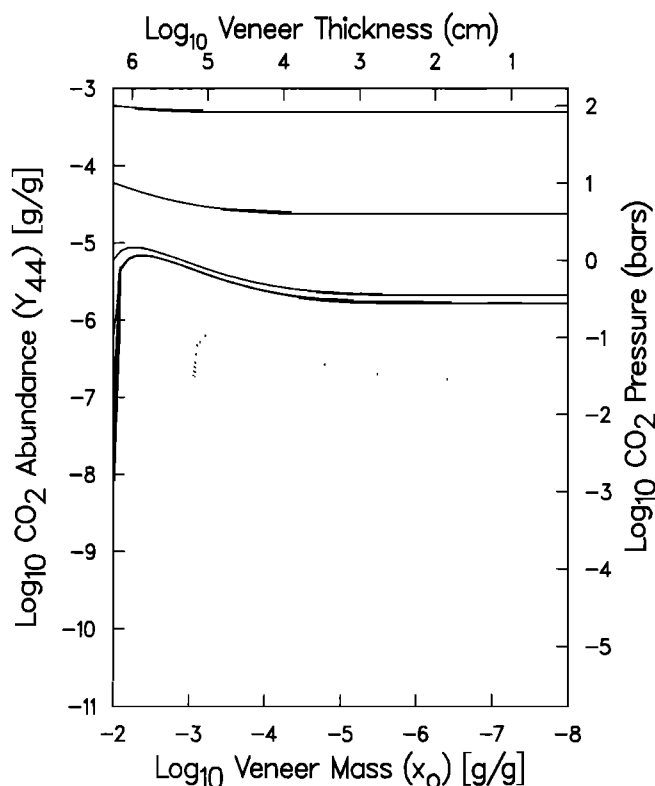


Fig. 6a. Evolutionary trajectories for inert, noncondensing CO_2 atmospheres under asteroidal bombardment (here $f = 0.7$, $y_{44} = 0.01$, and $q = 1.8$). This is a slightly accumulative case (i.e., it would plot above the relevant curve on Figure 5). Time goes to the right. Initial atmospheres range from 1 mbar to 100 bars by factors of 10. Masses of asteroids equivalent to veneers of $x_0 = 0.01$ (solid curves) and $x_0 = 0.001$ (dots) strike the planet. In this example initially thick atmospheres are little impacted, while initially thin ones equilibrate with the veneer.

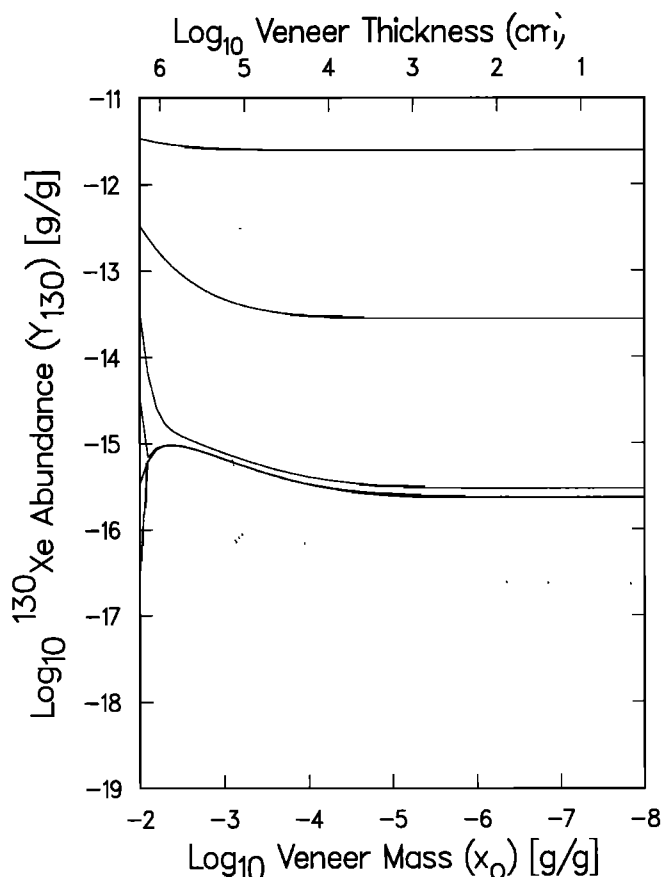


Fig. 6b. Same as 6a but for ^{130}Xe . An initial Martian $^{130}\text{Xe}/\text{CO}_2$ ratio is assumed.

the final ^{130}Xe abundances (i.e., Y_{130} as $x \rightarrow 0$) for the six initial ^{130}Xe inventories in Figure 6b for $10^{-7} < x_0 < 0.01$. The final states of the models in Figure 6b correspond to the points in Figure 6c at $x_0 = 0.01$ and $x_0 = 0.001$. Also shown in Figure 6c is the division between isotopically acceptable and unacceptable cases, where the isotopically acceptable cases are those above the dotted line (it is assumed that Martian atmospheric xenon can be no more than 20% chondritic). Figure 6c shows that, even when 70% of the impactors contribute to the atmosphere ($f = 0.7$), impacts can effect a great deal of erosion before their compositional fingerprint becomes visible. In this particular example Xe is reduced by a factor 20 before the erosive agents can be identified. On the other hand, this isn't enough to plausibly reproduce Mars; Xe should be reduced by at least a factor 70. This requires a more erosive regime, i.e., a lower value of f or y .

The case described in Figures 6a-c is slightly accumulative. A small change in parameters leads to a qualitatively different atmospheric evolution. Figure 7a for CO_2 evolution is identical to Figure 6a except that $f = 0.65$ rather than $f = 0.7$. Figure 7a shows that unless the atmosphere is initially thick (relative to the veneer), impacts remove the atmosphere, leaving a barren planet. Figure 7b, the analog to Figure 6c with $f = 0.65$, shows that if the veneer has just the right thickness a nice match to Mars can be achieved. Consider particularly the third curve from the top, which describes a 1-bar CO_2 atmosphere that is almost, but not quite, expelled by a 1 km veneer. Here Xe is reduced by nearly a factor of 100 before the chondritic

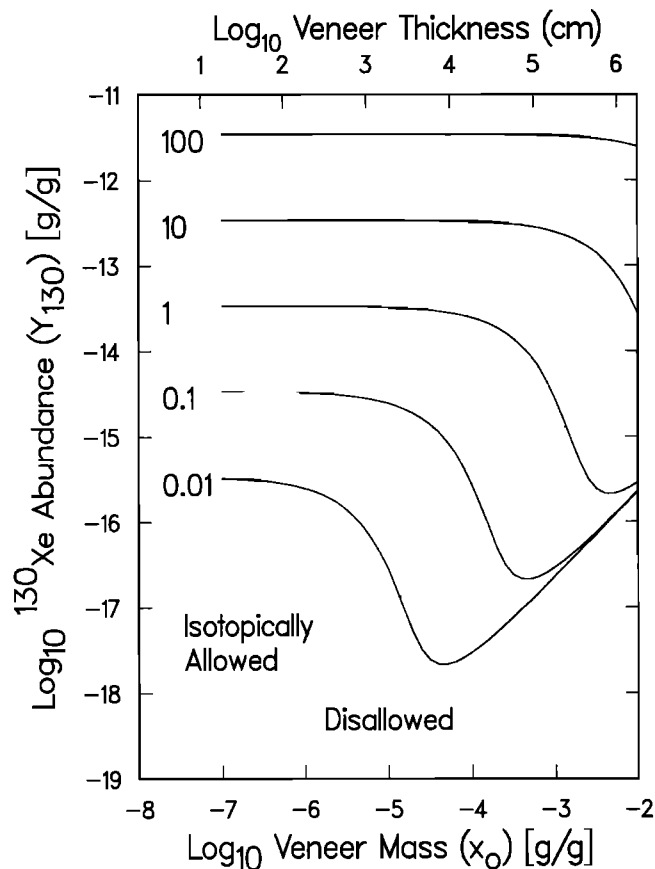


Fig. 6c. Final Martian ^{130}Xe inventories for the same six initial CO_2 atmospheres as a function total veneer mass x_0 . These are not evolutionary trajectories. The solid curves are marked by the original CO_2 atmosphere (bars). The dotted line indicates when atmospheric Xe has become 20% asteroidal. A "Mars" that plots below this line is isotopically disallowed.

pollutants become evident. Of course such fine tuning of the veneer thickness amounts to postulating special conditions. This is generally true for pure impact erosion models applied to Mars: it is possible to pick parameters that tell a good story, but the story is improbable. Only a narrow range of parameters will give both a thin atmosphere and one that looks unlike the impactors. In a plot like Figure 7b the measure of improbability is the narrow range of veneers (i.e., x_0) for which a given set of impactor parameters (q, y, f) can be made to work. A fair estimate of the probability that a thin atmosphere like Mars' would arise by impact erosion alone is no more than a few percent, although it might be a lot less. It may not be so unlikely that the existence of one thin atmosphere in the solar system would be an astonishment.

With $q = 1.8$, values of f larger than 0.7 produce rather thick atmospheres for Mars. At $f = 0.85$, closer to the preferred value for Mars according to Figure 3 and Table 1, there is little net erosion (Figure 8, analogous to Figures 6c and 7b). Carbon dioxide increases for all six initial atmospheres, with minimum accumulations of ~ 15 bars at $x_0 = 0.01$ and ~ 1.5 bars at $x_0 = 0.001$. It should be emphasized that the interesting values of f are sensitive to the specific values of y_{44} and q adopted; to give a particular example, models with $f = 0.45$ and $y_{44} = 0.03$ are quite similar to the models described above with $f = 0.7$ and

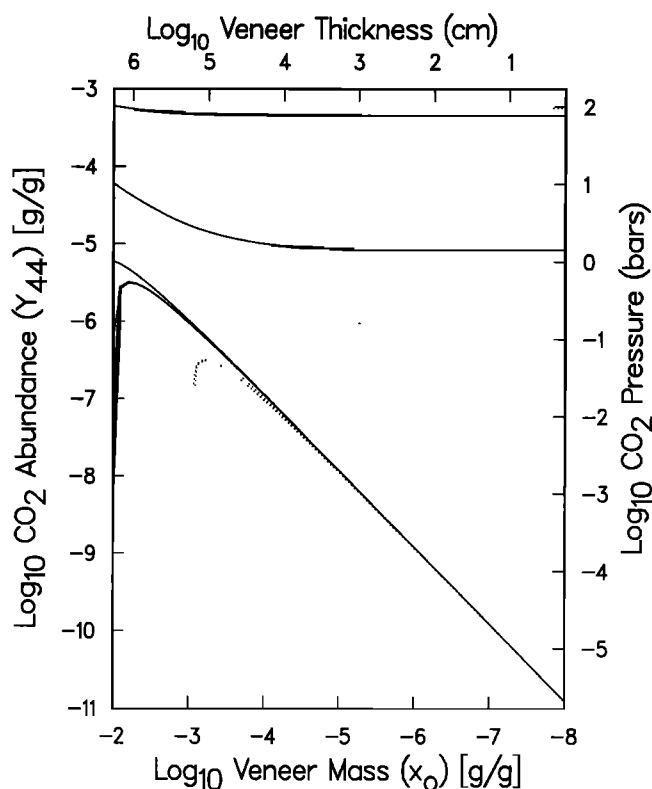


Fig. 7a. Like Figure 6b but for asteroids with $f = 0.65$. This is a slightly erosive case. Initially thick atmospheres are little impacted, while initially thin ones eventually vanish.

$y_{44} = 0.01$. A factor of 2 in y or f is obviously straining the limitations of the simple model of atmospheric ecology I've developed here. Still, my best guess of q, y, f for asteroids would place Mars among the planets that should develop thick atmospheres.

In summary, when planet-atmosphere exchanges are ignored, the more probable successful models yield either a Mars with a thick atmosphere or a Mars with no atmosphere at all. Other processes are probably required in addition to impact erosion. A possible way to retain an unpolluted remnant of a primordial Martian atmosphere is cold storage in the regolith. Even a strict atmophile can be caught underground in a porous regolith, especially if deep or subsequently sealed off by water ice. A regolith of depth z and porosity ϕ has room enough to harbor as much as a fraction $\phi(e^{z/H} - 1)$ of the atmosphere, where H is the atmospheric scale height. As a specific example, if $\phi = 0.05$ and $z = 10$ km, as much as 6% of the atmosphere could be underground. Because the atmosphere must compete with groundwater and ground ice for pore space, and since the upper parts of the regolith exchange quickly with the atmosphere and so make poor spaces for long-term storage, the actual fraction of the original atmosphere to survive the late bombardment underground will be less, perhaps of order 0.1-1%. Some of the ancient underground atmosphere would have been released relatively late in Martian history when water was mobilized, in a kind of outgassing.

Comets. Comets are more volatile-rich than asteroids but their volatiles are harder to capture. Assume a power law mass distribution governed by $q = 1.7$, a potential cometary CO_2 content of 20% by mass ($y_{44} = 0.2$), and a proportionately larger ^{130}Xe content of $y_{130} = 3 \times 10^{-11}$ g/g. These imaginary comets are so atmophile-rich that

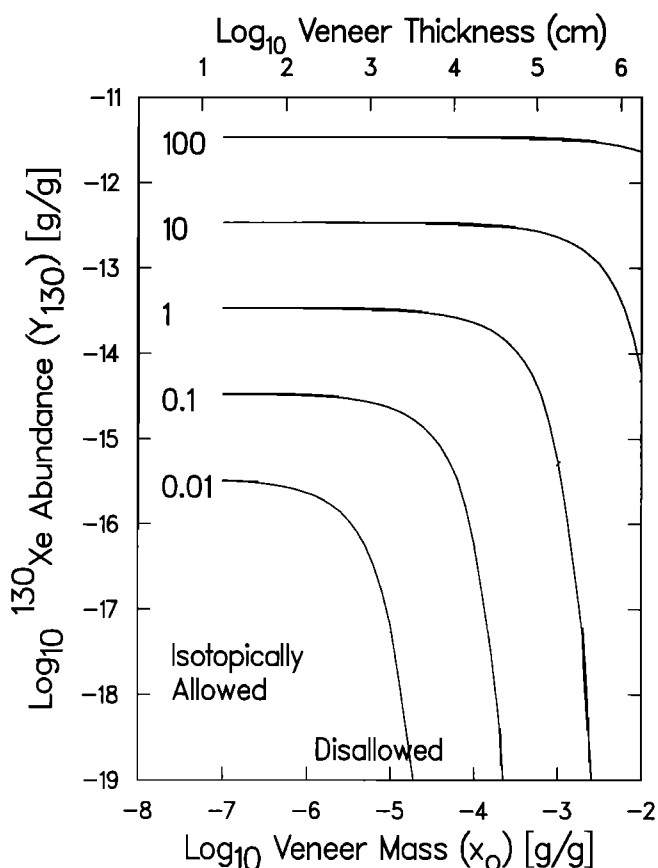


Fig. 7b. The analog to Figure 6c for $f = 0.65$. It is possible to choose parameters such that Xe is reduced a hundred fold while retaining its isotopic purity (e.g., the third curve from the top), but the veneer must have just the right thickness. If the veneer is slightly too thin too little Xe is eroded, and if it is slightly too thick nothing at all remains.

it is difficult not to accumulate a thick atmosphere if they are at all common. For the above parameters, Mars begins to accumulate a cometary atmosphere at about $f = 0.03$. Atmospheric evolutionary tracks for $f = 0.03$ and 0.02 are quite similar to the asteroidal tracks for $f = 0.7$ (Figure 6) and 0.65 , respectively. However, unlike asteroids, it is not clear whether the cometary flux can be constrained by the xenon isotopes. The xenon composition of comets is of course wholly unknown. On present knowledge the most reasonable isotopic composition would be solar; if so, the same isotopic constraints would limit comets as limit asteroids. But there remains open the tantalizing possibility that the distinctive noble gas composition of the terrestrial and Martian atmospheres is essentially that of comets, as has been suggested by Owen *et al.* [1991]. If so, a wider range of parameter space (in the region labeled "isotopically disallowed" on Figures 6 and 7) can produce acceptable cometary models than can produce successful carbonaceous chondritic models. In particular, models in which a thin atmosphere reaches compositional and volumetric equilibrium with a rain of impactors are possible. The requirement that impact erosion expel more than 99% of the atmosphere limits the acceptable range of f for a given q, y pair (and ensures that relatively little CO_2 remain), but given the right range of f , a fairly wide range of veneer thicknesses could work in the absence of the isotopic constraint.

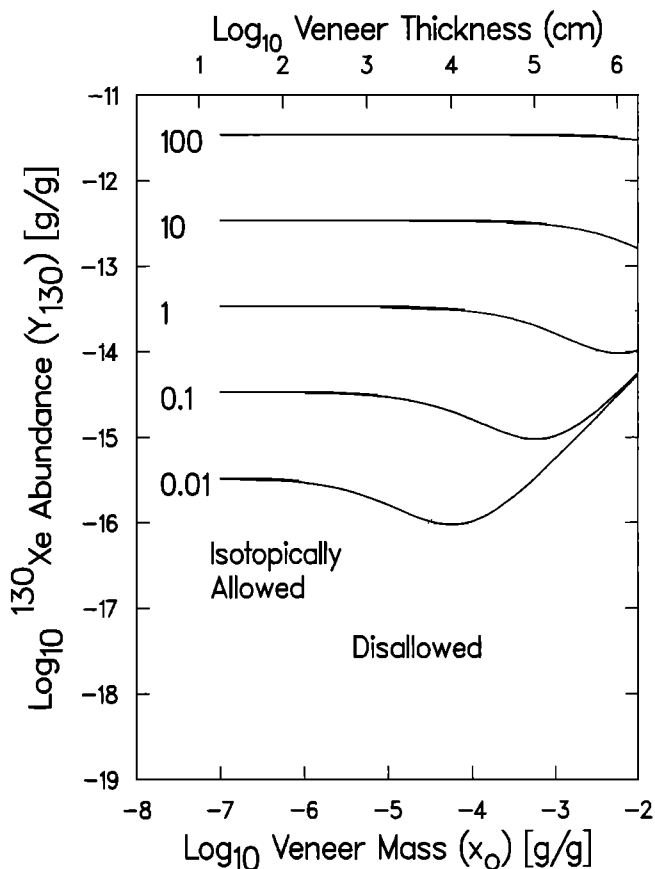


Fig. 8. Like Figures 6c and 7b but for asteroids with $f = 0.85$. This is a strongly accumulative case. Little net erosion occurs. These give nice thick CO_2 atmospheres in the end but are otherwise inappropriate for Mars, since far too much Xe remains.

IMPACT EROSION IN A CONSTANT BACKGROUND ATMOSPHERE

In the preceding section the atmosphere was treated as a mixture of inert gases enveloping an inert planet. An alternative that is relatively easy to treat is impact erosion acting on a constant background atmosphere. This could occur if the major atmospheric constituent had its primary reservoir elsewhere than the atmosphere. As noted above, a possible example would be CO_2 on Mars today if most Martian CO_2 were bound up as adsorbate or locked up in carbonate and if its atmospheric pressure is indeed held at the triple point of water. Over long periods of time CO_2 in Earth's atmosphere is geochemically controlled, so that it is not unreasonable to speculate that atmospheric CO_2 might have been geochemically controlled on early Mars, and that CO_2 was the major atmospheric constituent [Pollack *et al.*, 1987]. Another possibility is that atmospheric CO_2 on Mars may have been controlled by vapor pressure over a large surface reservoir of solid or even liquid CO_2 (a liberal reading of Kasting, [1991]).

With Y held constant, (17) is a first-order linear differential equation for $Y_2(x)$. It can be solved in closed form for $q = 1.5$. Figure 9a shows Xe evolutionary tracks for two veneers ($x_0 = 0.01$ and $x_0 = 0.001$) in background atmospheres of 8 mbar, 200 mbar, and 5 bars. The thinnest is representative of the modern Martian atmosphere, and the thickest is representative of the warm, wet early Mars [Cess *et al.*, 1980; Pollack *et al.*, 1987]. The middle case is a rough

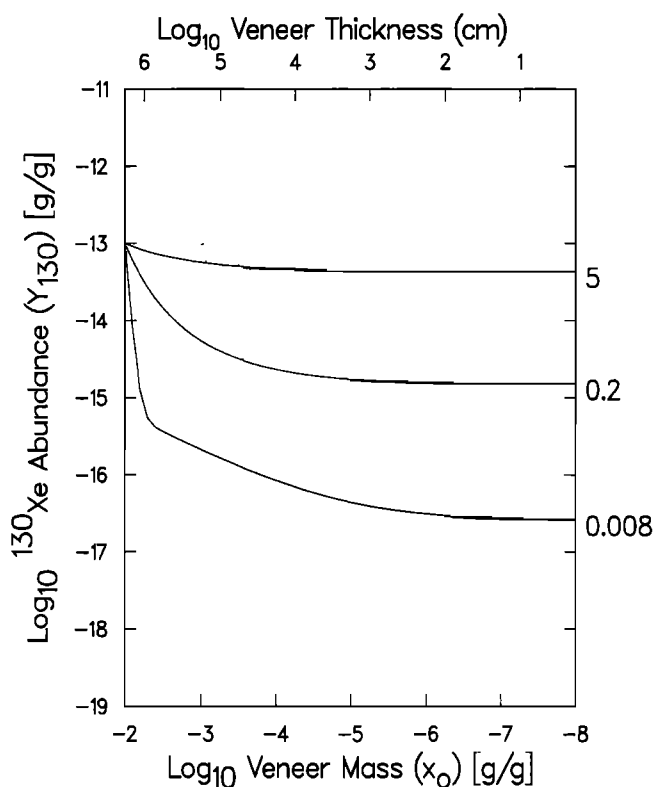


Fig. 9a. Xenon evolutionary tracks for asteroidal impact erosion acting on an atmosphere the main constituent (i.e., CO_2) of which is geochemically maintained at a constant pressure. Time goes to the right. The background CO_2 atmospheres are 8 mbar, 200 mbar, and 5 bars. The veneers are $x_0 = 0.01$ and $x_0 = 0.001$, and the impactors are characterized by $f = 0.5$, $q = 1.5$, and $y_{130} = 1.4 \times 10^{-12}$ g/g. For a given background atmosphere, Xe decays exponentially to a compositional equilibrium with the impactors.

upper limit on the amount of CO_2 that can be adsorbed in the regolith [Fanale and Jakosky, 1982] and thus may be viewed as an upper limit on "atmospheric" CO_2 on Mars today. In all six cases the initial planetary ^{130}Xe inventory is arbitrarily taken as 1×10^{-13} g/g. The impactors are described by $f = 0.5$, $q = 1.5$, and $y_{130} = 1.4 \times 10^{-12}$ g/g. This case may be illustrative of the late heavy bombardment. What is seen in Figure 9a is a rapid initial loss of xenon from the thinner atmospheres, followed by a more gradual equilibration with the impactors. For higher values of q the initial xenon loss is faster, and equilibration is quicker. In this particular example the thicker atmosphere is little affected by impact erosion, which is primarily a consequence of the low value of q used.

The composition of the atmosphere relaxes exponentially to the composition of the impactors. This is illustrated by Figure 9b, which is analogous to Figure 6c. Isotopic equilibration occurs more slowly than elemental equilibration, with a delay determined by the fraction of the initial Xe endowment that remains.

For $q \neq 1.5$, (17) is solved numerically. An illustrative asteroidal case ($f = 0.85$, $q = 1.8$, and $y_{130} = 1.4 \times 10^{-12}$) is the subject of Figure 10. It assumes an initial ^{130}Xe inventory of 1×10^{-13} g/g, and three geochemically maintained CO_2 atmospheres of 5 bars, 200 and 8 mbars. With the three curves shifted to the left by a factor five in incident veneer mass, Figure 10 can double as an illustration of a

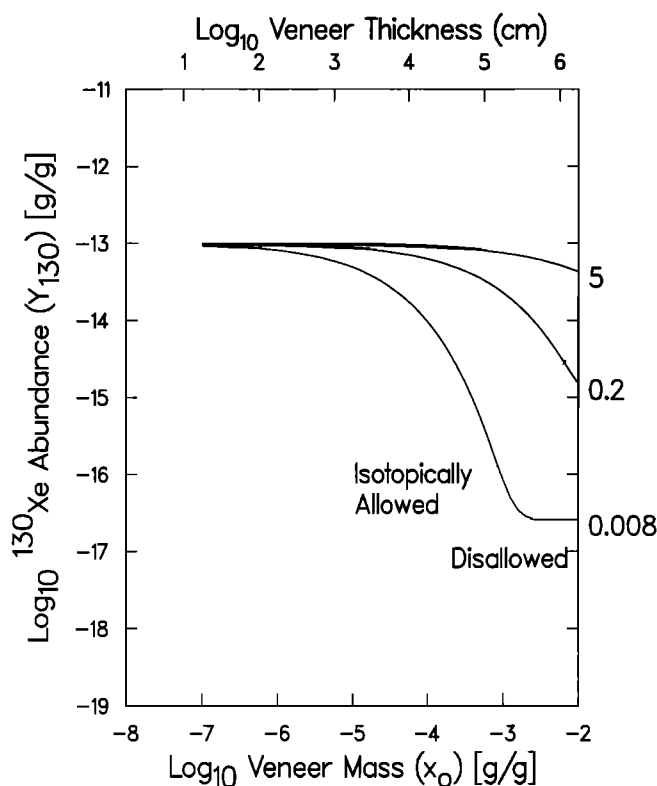


Fig. 9b. Final Martian ^{130}Xe inventories as a function total veneer mass x_0 (analogous Figure 6c etc.). These are not evolutionary trajectories. The gentler exponential decay illustrated in Figure 9a leaves more scope for erosive histories that greatly deplete Martian xenon yet leave the original isotopes little changed. Here an 8-mbar CO_2 atmosphere and a 1-km veneer perform admirably.

short- period "cometary" veneer with $f = 0.1$, $q = 1.7$, and $y_{130} = 5 \times 10^{-11}$ (i.e., a cometary veneer of $x_0 = 2 \times 10^{-4}$ is equivalent to an asteroidal veneer of $x_0 = 1 \times 10^{-3}$). In particular, the final steady states are the same; the higher assumed cometary Xe content is almost exactly balanced by the higher impact erosion rate. The atmosphere equilibrates more quickly with comets simply because cometary impacts are more erosive than asteroids. Both cometary and asteroidal impact predict substantial Xe losses for reasonably thin veneers even for a ~ 1 bar CO_2 atmosphere.

Perhaps the most interesting observation to be made in respect to Figure 10 is that the constant background atmosphere appears to moderate the effects of impact erosion, in that it is easier to build a planet that is highly deficient but not wholly bereft of xenon. The thick atmosphere protects exposed atmophiles like xenon. This is possible because the geochemically-maintained CO_2 atmosphere has been decoupled from impacts. In the previous section (c.f., Figures 6-8) where CO_2 is controlled by impacts it is CO_2 , not Xe, that determines whether Xe escapes or is retained. Thus what is essentially an erosive regime for Xe can be survived by a small but nonzero amount. Nevertheless the isotopic constraint continues to severely limit the likelihood of modern Martian Xe arising by impact erosion of an initially distinctive Martian Xe.

The prospects for isotopically Martian comets are also improved by the presence of a background geochemically patrolled atmosphere. It is relatively easy to choose reasonable Martian parameters for a cometary source that

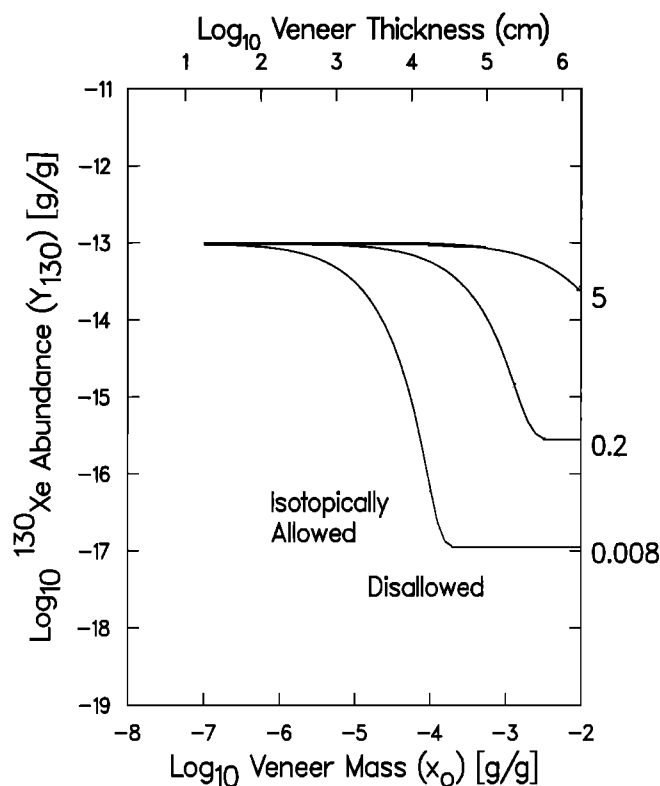


Fig. 10. Like Figure 9b, but for asteroids with $f = 0.85$ and $q = 1.8$. An essentially identical plot, only shifted to the left by a factor five in veneer mass, results for comets with $f = 0.1$ and $q = 1.7$. Because q is higher, Figure 10 describes a more erosive regime than Figure 9. Here a 200-mbar CO_2 atmosphere and a 1-km veneer do well. However, for a given CO_2 atmosphere only a narrow range of veneer masses gives both greatly depleted Xe and acceptable isotopes. If cometary xenon is isotopically solar, the same restrictions hold against comets as hold against asteroids. If, on the other hand, cometary Xe is isotopically fractionated like the atmosphere, a wide range of equilibrium solutions in the lower right hand side of Figure 10 become acceptable. Thus an isotopically correct cometary veneer impacting a moderately thick geochemically regulated CO_2 atmosphere offers a wide target for an impact erosion history of Mars.

will leave a small but nonzero xenon inventory that looks isotopically like the impactors themselves. Whether real comets are characterized by profoundly mass fractionated xenon is an open question, but if they should prove so, Mars becomes a powerful argument in favor comets as the dominant source of planetary volatiles.

SOME CONCLUSIONS

I find that if impact erosion of a planetary atmosphere is important, the remnant atmosphere is likely to fall into one of two qualitatively distinct categories. Neither category can readily account for the present Martian atmosphere. Either cometary xenon has an unexpected isotopic composition or factors other than impacts, e.g., outgassing or recycling, are needed.

In one category impact erosion is catastrophic, in the sense that if the atmosphere is subject to substantial erosion, the atmosphere will most likely be wholly lost. This can happen if all atmospheric constituents have their primary reservoirs in the atmosphere (so that impacts are the largest sink and source) and if escape is dominated by the

"tangent-plane" process, which, as suggested by *Melosh and Vickery* [1989], should be the case if impact velocities are often greater than about twice the escape velocity. The tangent-plane process is catastrophic because the smallest impactor capable of effecting escape is proportional to the mass of the atmosphere, so that as the atmosphere shrinks, impacts that can erode it further become smaller and more numerous. Thus survival of a thin atmosphere (i.e., Mars) is inherently unlikely.

In the other category the major atmospheric constituent is externally controlled by a larger crustal or mantle reservoir, as may have been and may still be the case for CO₂ on Mars. Impact erosion acting on this kind of atmosphere can easily reduce strict atmophiles like nitrogen and the noble gases to very low equilibrium abundances without causing them to vanish altogether. However, the elemental and isotopic abundances of the strict atmophiles eventually come to resemble the impactors. Therefore either Mars is unlikely, with impact erosion arrested before the atmosphere equilibrates with the impactors, or the noble gases in the impactors resembled the noble gases in the Martian atmosphere rather than modern meteorites. The obvious candidate for an unknown, planet-like source is to site the noble gases in cometary ices. A common cometary origin would neatly account for Earth and Mars [*Owen et al.*, 1991], but otherwise cometary origin is problematic; the desired composition, marked by a supersolar Kr/Xe ratio and profound xenon isotopic fractionation with respect to solar, does not align well with one's naive expectation that cometary ice would either be of solar composition, or be deficient in the lighter elements. In any event this hypothesis is subject to telling experimental tests.

These qualitative conclusions are not affected by reasonable uncertainties regarding the efficiency of the processes. Various complicating details could make tangent-plane escape much less efficient than estimated by Melosh and Vickery without altering the qualitative conclusion that either Mars is unlikely or comets look like air.

How unlikely? A quantitative estimate of how unlikely it would be to have just enough impact erosion to reduce the atmosphere a hundred or a thousandfold yet not so much that either the atmosphere is eliminated or the impactors pollute the remnant is difficult. There are at least four relevant parameters (f , the fraction of impacts with impact velocities below the threshold for atmospheric cratering; y , the average atmophile content of the impactors; q , the power law exponent that describes the mass spectrum of the impactors; and x_0 , the mass of the late accreting veneer) that need to be constrained. For a given (q, y) pair only limited ranges of both f and x_0 can yield the desired history; thus the a priori likelihood of a thin, unpolluted atmosphere resulting from impact erosion is the product of two small numbers. I estimate this as no more than a few percent. A more accurate measure requires better knowledge of the probable ranges of the four parameters.

In my opinion the best story is that the primordial Martian atmosphere was indeed completely stripped by impacts, and this very early, most likely in the first 100 m.y. The present Martian atmosphere would have been a small part of the primordial atmosphere safely stored deep in the regolith, possibly trapped in ice. The deep regolith is much less subject to impact erosion than is the atmosphere. The stored gases would be released later over a longer time

scale, perhaps through low-grade volcanism associated with the mobilization of water. The present Martian atmosphere would therefore be an accurate measure of the present total atmophile inventories, most likely including nitrogen, and possibly including CO₂, as well.

The primary advantages of the "hidden atmophile" hypothesis are that (1) it is consistent with catastrophic tangent-plane atmospheric cratering, which is the most likely mode of impact erosion for Mars; (2) indigenous xenon avoids contamination by late impactors; (3) non-radiogenic noble gases are removed very quickly, on a time scale determined by the late phases of planetary accretion, and thus are consistent with the observed retention of radiogenic ¹²⁹Xe; and (4) the retention of about 0.1–1% of the initial nonradiogenic noble gases is consistent with the capacity of the regolith to harbor foreign atmophiles. This hypothesis places fewer constraints on condensible volatiles like water or CO₂, especially those that can be stored deep in the regolith. Water, in particular, is almost certainly retained.

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K. Zahnle, NASA Ames Research Center, MS 245-3, Moffett Field, CA 94035-1000.

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