escape would produce a fractionation of $^{15}N/^{14}N$ larger that observed; an early, thicker CO_2 atmosphere could mitigate the N loss and produce the observed fractionation. The total amount of CO_2 lost over geologic time is probably of the order of tens of millibars rather than a substantial fraction of a bar. The total loss from solar-wind-induced sputtering and photochemical escape, therefore, does not seem to be able to explain the loss of a putative thick, early atmosphere without requiring formation of extensive surface carbonate deposits.

POSSIBLE SOLUTIONS TO THE PROBLEM OF CHANNEL FORMATION ON EARLY MARS. J. F. Kasting, Department of Geosciences, 211 Deike, Pennsylvania State University, University Park PA 16802, USA.

A warm climate on early Mars would provide a natural, although not unique, explanation for the presence of fluvial networks on the ancient, heavily cratered terrains. Explaining how the climate could have been kept warm, however, is not easy. The idea that the global average surface temperature, T_s, could have been kept warm by a dense, CO_2 atmosphere supplied by volcanism or impacts [1,2] is no longer viable. It has been shown that CO_2 cloud formation should have kept T_s well below freezing until ~2 b.y. ago, when the Sun had brightened to at least 86% of its present value [3] (Fig. 1). Warm equatorial regions on an otherwise cold planet seem unlikely because atmospheric CO_2 would probably condense out at the poles. Warming by impact-produced dust in the atmosphere seems unlikely because the amount of warming expected for silicate dust particles is relatively small [4]. Greenhouse warming by high-altitude CO_2 ice clouds seems unlikely because such clouds are poor absorbers of infrared radiation at most wavelengths [5]. Warming by

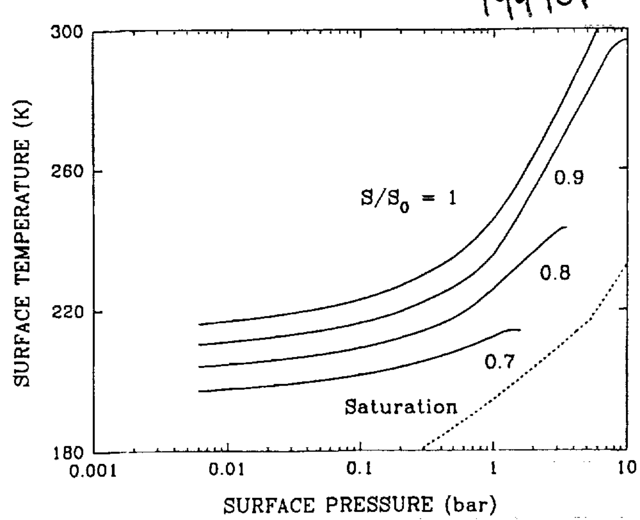

Fig. 1. Mean global surface temperature on Mars as a function of atmospheric CO_2 partial pressure. S/S_0 represents the magnitude of the solar luminosity compared to its present value. Solutions with mean surface temperature >273 K are found only for S/S_0 > 0.86. The dashed curve is the saturation vapor pressure curve for CO_2. (From [3].)

atmospheric NH_3 [6] seems unlikely because NH_3 is readily photodissociated [7] and because N may have been in short supply as a consequence of impact erosion [8] and the high solubility of NH_3. A brighter, mass-losing young Sun [9] seems unlikely because stellar winds of the required strength have not been observed on other solar-type stars. In short, most of the explanations for a warm martian paleoclimate that have been proposed in the past seem unlikely.

One possibility that seems feasible from a radiative/photochemical standpoint is that CH_4 and associated hydrocarbon gases and particles contributed substantially to the greenhouse effect on early Mars. Methane is photochemically more stable than NH_3 and the gases and particles that can be formed from it are all good absorbers of infrared radiation. The idea of a CH_4-rich martian paleoatmosphere was suggested a long time ago [10] but has fallen out of favor because of perceived difficulties in maintaining a CH_4-rich atmosphere. In particular, it is not obvious where the CH_4 might come from, since volcanic gases (on Earth, at least) contain very little CH_4. This difficulty could be largely overcome if early Mars was inhabited by microorganisms. Then, methanogenic bacteria living in sediments could presumably have supplied CH_4 to the atmosphere in copious quantities.

Thus, if I were a betting scientist, I would wager that either early Mars was inhabited, or the martian channels were formed by recycling of subsurface water under a cold climate, as proposed by Clifford [11] and others.

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CORE FORMATION, WET EARLY MANTLE, AND H_2O DEGASSING ON EARLY MARS. K. Kuramoto and T. Matsui, Department of Earth and Planetary Physics, University of Tokyo, Bunkyo-ku, Tokyo 113, Japan.

Introduction: Geophysical and geochemical observations strongly suggest a "hot origin of Mars," i.e., the early formation of both the core and the crust-mantle system either during or just after planetary accretion [1]. To consider the behavior of H_2O in the planetary interior it is specifically important to determine by what mechanism the planet is heated enough to cause melting. For Mars, the main heat source is probably accretional heating. Because Mars is small, the accretion energy needs to be effectively retained in its interior. Therefore, we first discuss the three candidates of heat retention mechanism: (1) the blanketing effect of the primordial H_2-He atmosphere, (2) the blanketing effect of the impact-induced H_2O-CO_2 atmosphere, and (3) the higher deposition efficiency of impact energy due to larger impacts. We conclude that (3) the is the most plausible mechanism for Mars. Then, we discuss its possible consequence on how wet the early martian mantle was.

Early Thermal History: If Mars grows under the presence of solar nebula gas, the primordial H_2-He-type atmosphere may surround a growing planet [2]. By assuming the adiabatic temperature gradient in the primordial atmosphere, the maximum surface temperature T_{max} resulting from the blanketing effect of this type of atmosphere can be estimated. This is given by

$$T_{max} \approx \frac{\gamma - 1}{\gamma} \frac{GM\mu m_H}{k_B R}$$

where G is the gravitational constant, M and R are the mass and radius of the planet, μ is the mean molecular weight of atmospheric gas, k_B is the Boltzmann's constant, m_H is the mass of a hydrogen atom, and γ is the ratio of specific heats of atmospheric gas at constant pressure and volume. Taking $M = 6 \times 10^{23}$ kg, R = 3400 km, $\gamma = 1.4$, and $\mu = 2$, T_{max} is about 830 K. This estimated value is too low to cause the melting of planetary material.

On the other hand, if Mars grows mostly after the dissipation of solar nebula gas, the impact-induced H_2O-CO_2 atmosphere may be formed [3]. In the case of such an atmosphere, the accretion energy flux ≥ 200 W/m^2 is likely to be required at Mars' orbit for sustaining the surface temperature above 1500 K by the blanketing effect [4]. The mean accretion energy flux F_{acc} may be estimated by

$$F_{acc} = \frac{M}{\tau_{acc}} \left(\frac{GM}{R} + \frac{\upsilon_\infty^2}{2} \right) / 4\pi R^2$$

where τ_{acc} is the accretion time and υ_∞ is the random velocity of the planetesimal at infinity. If $\tau_{acc} = 10^8$ yr for Mars [5], F_{acc} are 15, 81, and 277 W/m^2, respectively, when $\upsilon_\infty = 0$, 10, and 20 km/s. Thus, the accretion energy flux ≥ 200 W/m^2 is not available unless the random velocity is significantly large. Although such a large random velocity may be possible, impacts with such a high velocity are likely to erode the atmosphere [6]. We may therefore conclude that the blanketing effect of neither the primary H_2-He type nor the impact-induced H_2O-CO_2 atmosphere is plausible to explain hot early Mars.

Recently, [7] proposed that the formation of large magma ponds due to massive impacts plays a key role in the core formation of terrestrial planets, if impact velocities were larger than 7.25 km/s. They assume that half the impact energy is imparted to the kinetic and internal energies of the target during the compression stage. However, numerical simulation of crater formation suggests that more than 80% of the impact energy is imparted to the target [8]. Therefore, we reestimate the volume of impact melt by combining Tonks and Melosh's approach and the energy partitioning estimated by [8]. The reestimated volume is about 3× larger than that by [7] under the given impact velocity and mass of the planetesimal. The impact of a planetesimal larger than 10^{18} kg may form a sufficiently large magma pond on Mars. An iron blob formed in such a magma pond is large enough to sink through the relatively cold interior (assuming viscosity = 10^{21} Pa s) to the center within 10^8 yr. Such planetesimals with mass $\geq 10^{18}$ kg and impact velocity ≥ 7.25 km/s are physically plausible [5]. Thus, the formation of such magma ponds by large impacts may be the most plausible mechanism to explain the early formation of the core in Mars.

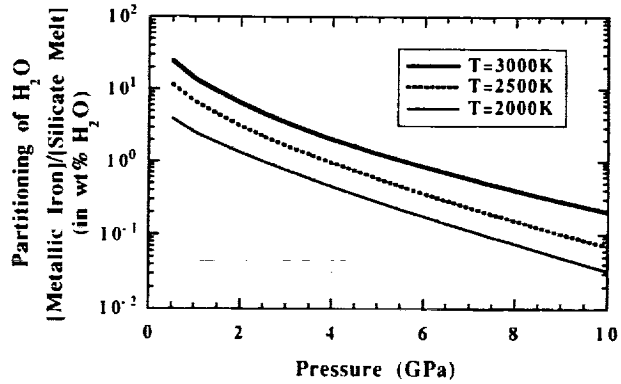

Fig. 1. H_2O partitioning between the silicate melt and liquid metallic Fe. Hydrogen concentration in metallic Fe is shown as a value converted to equivalent H_2O concentration. H_2O fugacity in silicate melt is calculated from the result by [9]. H_2 fugacity in liquid Fe is calculated from the model by [10]. The ratio of fugacities between H_2O and H_2 is given by the quartz-iron-fayalite O buffer and dissociation equilibrium of H_2O. H_2O concentration in metallic Fe is assumed to be 1.0 wt% in this calculation. Note that under pressure lower than several GPa (correponds to several 100 km depth in Mars), H_2O is distributed in silicate melt more than metallic Fe.

Differentiation in a Magma Pond: Next let us consider the process of chemical differentiation in a magma pond. Just after the formation of a magma pond, Fe particles are settled to form a large Fe blob [7]. The Fe particles are probably in chemical equilibrium with the surrounding silicate melt during segregation. On the other hand, the large Fe blob is not in chemical equilibrium with the surrounding mantle during further sinking. A certain fraction of H_2O contained in the planetesimal is probably incorporated into such a magma pond because H_2O can dissolve into silicate melt. Water may be partitioned between silicate melt and metallic iron in the magma pond. We can estimate this partitioning using a thermodynamic model. As shown in Fig. 1, the partitioning of H_2O is dependent on the pressure. This means that the depth of magma ponds may affect the behavior of H_2O in the interior of a growing Mars.

Early Formation of Hydrosphere: If the typical depth of magma ponds is smaller than several 100 km, H_2O is mainly retained in silicate melt as shown in Fig. 1. In this case, the protomartian mantle may be wet. The possible existence of H_2O in the protomartian mantle may have a profound influence on the formation of the surface hydrosphere on Mars. Water was able to degas extensively from the wet protomantle associated with early crustal formation. This may become a prime source of surface H_2O suggested from geologic flow features on early Mars.

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BRINY LAKES ON EARLY MARS? TERRESTRIAL INTRACRATER PLAYAS AND MARTIAN CANDIDATES.

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Recently, salt-rich aqueous solutions have been invoked in the preterrestrial alteration of the Nakhla [1] and Lafayette [2] SNC meteorites. The findings substantiate the long-standing suspicion that salts are abundant on Mars [3] and, more importantly, that brines have played a significant role in martian hydrogeological history. Adding to the growing body of evidence, I report here on the identification of several unusual intracrater high-albedo features in the ancient cratered highlands of Mars, which I interpret as possible saline playas, or salt pans.

The formations appear in Mariner 9 and Viking Orbiter images as relatively bright, low-relief, irregular or ring-shaped mottled patches on the floor of certain impact craters. Occasionally exhibiting a well-defined bright "core" usually 5–10 km in length, the features may stretch over several tens of kilometers as subdued, bathtub-ring-like formations. Most are located at equatorial to mid latitudes, with a distinct clustering of the core-bearing features in Terra Tyrrhena (e.g., at 11.5°S,290.5°W, 19.2°S,291.75°W, or 19.6°S,300.5°W). These features are unlikely to be merely eolian deposits because their albedo patterns and general outlines are unlike those of common martian or terrestrial eolian formations. Moreover, they often occur in portions of crater floors that would not be expected, based on the indications of nearby wind markers, to exhibit eolian deposits. The features are also unlikely to be CO_2 or H_2O frost cover because many of them lie at latitudes no higher than ±20 where such frosts would be thermodynamically unstable under present martian conditions. The features show no appreciable morphological change over time intervals spanning more than 2 martian years and, from preliminary cratering statistics, actually appear to be very ancient. On the basis of their distinct morphology, mode of occurrence, and stability through time, and also because there is strong, albeit indirect, evidence for the presence of salts [3] and for the past presence of ponding water on Mars [e.g., 4], I suggest instead that these features are plausibly evaporitic saline playas.

Geochemical measurements performed at the Viking Lander sites [5] and on SNC meteorites [1,2] indicate that salts are probably an important component of the martian regolith. Also, geomorphologic and climatic observations, in particular the identification of ancient aqueous sedimentary basins on Mars [4], suggest that saline playas may be expected to have formed. Until now, however, no large-scale salt pans have been identified on Mars, with the possible exception of the intracrater high-albedo, low-relief feature known as White Rock (8°S,335°W) [6]. While morphologically not identical to White Rock, the features reported here have similar attributes and may be of the same nature. All present many of the morphologic and setting characteristics that are associated with saline playas on Earth. (Note that none of these formations, including White Rock, are actually white, at least partly because of the ubiquitous presence of pigmented eolian fines that cover much of the martian surface.) These considerations prompted a reexamination of terrestrial examples of intracrater playas. Impact craters often constitute enclosed drainage basins that, when exposed to arid or semiarid environments, commonly develop playa units on their floors. Such playas are of particular value because their record of paleoenvironments was kept confined and is likely to have been well preserved. Examples of intracrater playas include those found within Meteor Crater (USA) and the Acraman, Connolly Basin, Henbury, and Wolf Creek Craters (Australia). At Wolf Creek, the center of the crater floor is occupied by a saline playa 450 m in diameter [7]. It is composed of porous gypsum and pitted with sink holes that expose underlying calcareous tuff. The origin of the gypsum in the crater's quartzitic setting is unclear. While Mason suggested eolian influx of Ca-rich feldspathic sands, McCall hypothesized intracrater hot spring activity [8]. A simpler explanation, however, common to many extracrater playas, might just involve discharges of briny groundwater accompanying secular variations in the level of the water table.

If the features reported here are indeed saline playas, implications for the evolution of the martian surface would be very important. They would, for instance, suggest that there were once relatively large intracrater briny lakes on Mars. It is not clear, however, whether the implied transient bodies of liquid water would have resulted from surface runoff, direct precipitation, and/or groundwater discharge. While the small-scale surface texture of terrestrial playas (at meter and submeter scales) are often diagnostic of their mode of formation and their subsequent history of modification, images of higher resolution than those provided by the Viking orbiters are needed before such information may be derived for the martian features. The possible existence of saline playas on Mars is tantalizing also because weathering by salts has been suggested as a significant process of geological alteration [9]. The preservation of secular ice in high-altitude Andean salars, under high-albedo, thermally insulating, and diffusion-inhibiting blankets of salts [10], further underscores the potential value of possible analogs on Mars. High-resolution images acquired by the Mars Observer Camera, along with targeted measurements by the Thermal Emission Spectrometer, offer a unique opportunity to test these ideas and to gain a better understanding of the intriguing features.

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