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WORKSHOP ON EARLY MARS: HOW WARM AND HOW WET?

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WORKSHOP ON

EARLY MARS: HOW WARM AND HOW WET?

Edited by

S. Squyres and J. Kasting

Held in
Breckenridge, Colorado

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Preface

This volume contains papers that have been accepted for presentation at the Workshop on Early Mars: How Warm and How Wet?, July 26–28, 1993, in Breckenridge, Colorado. The Program Committee consisted of S. Squyres (*Cornell University*) and J. Kasting (*Pennsylvania State University*).

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Abstracts

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EVOLUTION OF THE GLOBAL WATER CYCLE ON MARS: THE GEOLOGICAL EVIDENCE. V. R. Baker and V. C. Gulick, Department of Geosciences and Lunar and Planetary Laboratory, University of Arizona, Tucson AZ 84721, USA.

The geological evidence for active water cycling early in the history of Mars (Noachian geological system or heavy bombardment) consists almost exclusively of fluvial valley networks in the heavily cratered uplands of the planet [1–3]. It is commonly assumed that these landforms require explanation by atmospheric processes operating above the freezing point of water and at high pressure to allow rainfall and liquid surface runoff [4–6]. However, it has also been documented that nearly all valley networks probably formed by subsurface outflow and sapping erosion, involving groundwater outflow prior to surface-water flow [7–10]. The prolonged groundwater flow also requires extensive water cycling to maintain hydraulic gradients, but need this be via rainfall recharge, as in terrestrial environments [11,12]?

It is important to contrast the Noachian-age evidence of water cycling with that proposed for later epochs of martian history (postheavy bombardment or Hesperian and Amazonian geological systems). Valleys formed at these times [13,14] coincident with glaciation [15], periglacial mass movement [16,17], and ponded water on the northern plains [18,19]. The latter may have constituted transient oceans, fed by outflow channel discharges [20,21], that temporarily induced short-term phases of warm, wet atmospheric conditions [22]. Extensive glaciation in middle Amazonian time [23] required a short-term epoch of precipitation. Unequivocal geological evidence for precipitation does not occur for Noachian time.

The argument for a Noachian ocean on Mars is theoretical [24]; confirming geological landforms are not present. Moreover, there are strong arguments against the possibility of a CO_2 greenhouse warming of the atmosphere to temperatures above freezing [e.g., 25] during the heavy bombardment epoch coinciding with the faint young Sun [26]. Theoretical resolutions of this "paradox" include modified solar theory [27] or the development of a reducing atmosphere greenhouse [28,29].

The theoretical arguments for the early Mars warm, wet scenario (EMWS) is predicated upon far less compelling geological evidence than that for the episodic late Mars climatic change hypothesis [22], since only valley networks create the rationale. Crater degradation by pluvial action [30] is alternatively explained by backwasting and downwasting through mass movement [31,32], indicating a need for resolution of the conflicting interpretations. Without unequivocal evidence of glaciers, oceans, or other indicators of global pluvial processes, there is the possibility of an alternative case of nonpluvial water cycling to explain the valley networks.

Large quantities of groundwater can by cycled in the vicinity of subsurface thermal anomalies through hydrothermal circulation [33,34]. Possible thermal anomaly sources include igneous sills [35,36], volcanos [37], and impacts [38]. During the Noachian, planetary heat flow was much greater [39], yielding high near-surface temperatures [40]. However, these do not, by themselves, account for the prolonged circulation necessary to form valleys. Ratios of water volume circulated to excavated valley volume are estimated as low as 4 [41,42], but terrestrial analogs suggest values more on the order of 10^3 [43]. Localized hydrothermal circulation seems the only possibility for yielding the higher ratio.

Noachian valley formation by hydrothermal processes does not require the EMWS, but snowfall would certainly enhance the process of valley formation with recharge into hydrothermal warm spots. The distributed volcanism of the late Noachian intercrater plains is consistent with the formation of valleys by hydrothermal circulation, explaining the low drainage densities [44] and clustered distributions of valleys.

Mars erosional episodes are clustered in time [45]. The Noachian epoch was followed by Hesperian and Amazonian periods of valley formation [22]. Volcanism also became localized over the same period, eventually clustering at Elysium and Tharsis. Massive Hesperian and Amazonian outflows probably explain the depositional mantles of the northern plains [46,47]. The Hesperian/Amazonian ages of plains materials [48–50] are in accord with this model of post-Noachian change. The EMWS may well have been endogenetically induced by more uniform volcanism in space and time, while the cataclysmic episodic phases of later epochs of warmth on Mars reflect a subsequent localization of volcanism in time and space [22].

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52-91 NBS ONLY 177592

ANCIENT MARTIAN VALLEY GENESIS AND PALEOCLIMATIC INFERENCE: THE PRESENT AS A KEY TO THE PAST. G. R. Brakenridge, Surficial Processes Laboratory, Department of Geography, Dartmouth College, Hanover NH 03755, USA.

Understanding the origin of the relict fluvial landforms that dissect heavily cratered terrains on Mars is one of the clearest challenges to geomorphology to emerge in this century. The challenge is not being successfully met. Twenty years after the discovery of these landscapes we still do not know how the valleys formed or what they imply regarding paleoclimate. Geomorphology has not, to date, provided robust, remote-sensing-related analytical tools that are appropriate to the problem. It cannot because, despite the truism that "form follows function," landform genesis on Earth is not studied today by matching landform morphometries to genetic processes. Instead, the lithology, structure, and stratigraphy of the material underlying the landform is sought, and the local geological history may be investigated in order to understand the stage on which modern surficial processes are playing. Unfortunately, not much information other than morphometry has been assembled for the ancient martian valleys.

The problem is complicated by the relict nature of the landforms and the apparent need to infer genetic processes that are no longer occurring: the past must be reconstructed. To this end, geoscientists commonly use the principle of "uniformitarianism," wherein inferences regarding the past are based on the reality of the present. Consider the limitations that a comparative planetology approach presents, e.g., to a geologist from Mars attempting to understand the genesis of the now-relict Appalachian Mountains. Mars-Earth comparisons could not yield much insight. Understanding the passive margin Appalachians is critically dependent on a chain of insights regarding processes that derive from Earth's present. Terrestrial crustal plates move, and sea floor spreading, subduction, continental collision, and subsequent rifting occur. Inferring such processes would seem wildly speculative from a Mars perspective, and the genetic chain for creating relict mountains is indeed complex. However, we know that the Appalachians were in fact so created, and we gained such knowledge by extrapolation from the present (e.g., the Himalayas). When investigating the genesis of relict fluvial valleys on Mars, we could do the same.

For example, immediately inside the rim of the 120-km-diameter crater Cerruli (located at ~+30°, 340°), 1-km-wide, flat-floored and (narrower) v-shaped valleys debouch from apparent 5–10-km-wide collapse depressions and extend for distances of a few tens of kilometers downslope and toward the center of the crater [1]. None of the valley landforms examined are cratered; they may have formed quite recently or may still be active. Intervening preserved highland remnants may be composed of ice-rich sedimentary material [1], and the large collapse depressions exhibit margin-proximal parallel lineations suggestive of margin-derived sediment input.

The terrain bears little resemblance to fluvial landscapes on Earth, but is similar to ancient, fluvially dissected terrain in Aeolis Quadrangle [2]. The flat-floored valleys do crudely resemble terrestrial glaciated valleys, and parabolic viscous drag-flow lineation is expressed locally on their smooth-surfaced floors, suggesting wall-to-wall longitudinal transport of debris and/or ice. However, these valleys exhibit scalloped margins and no snowfields feed their headwaters. Instead, and at varying distances upslope from the abrupt amphitheater headwalls, shallow closed depressions are visible in Viking Orbiter stereopairs (204S18-21), and these may indicate embryonic collapse and possible future headward growth of the valleys.

Viking-based stereoscopy also demonstrates that low-gradient or flat plains separate steep-gradient, incised, v-shaped valley reaches. At the limit of resolution, narrow channels on these plains appear to connect the valley reaches, but the incised reaches are perhaps more akin to avalanche chutes than to terrestrial fluvial valleys.

There is evidence for at least two different valley genesis pathways on sloping terrain inside Cerruli's complex crater rim: (1) collapse, linking of collapsed areas by growing flat-floored valleys, and transport of valley floor material in the downvalley direction even as valley widths grow, at least slightly, by mass-wasting; and (2) carving of much narrower, sometimes en echelon valleys along steep hillsides, coupled with suggestive evidence of intervening flow in channels on relatively flat terrain.

I offer here the speculative genetic hypothesis that the flat-floored landforms represent episodically active, sediment-laden valley glaciers formed by localized geothermal melting of abundant interstitial ice (permafrost) in a fine-grained sedimentary terrain. Geothermal melting may also localize spring heads for the narrow, deep, high-gradient valleys, or the collapse process itself may result in the generation of decanted, relatively sediment-poor overland water flows (some local evidence of fluid overtopping of the localized depressions exists). Whatever the genetic mechanisms for the suite of valley landforms, perhaps the most interesting observation is simply their youth. In aggregate, the morphologies are similar to the ancient valley systems cited as evidence for a previously much denser atmosphere on Mars.

If even very local valley genesis occurs today and the landforms are similar to those dating from ca. 3.8 Ga, then global climatic cooling may not be the most appropriate shut-off mechanism for the ancient valleys. Other mechanisms include (1) a change in planetary volcanism style from mainly effusive plains volcanism to plumate eruptions [2], (2) changes in orbital parameters destabilizing ground ice at low latitudes but permitting stability at higher latitudes [3], or

(3) climatic warming, perhaps coeval to the termination of the Sun's T Tauri phase, and resulting poleward latitudinal shift of the zones of preserved ground ice.

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A CARBON DIOXIDE/METHANE GREENHOUSE ATMOSPHERE ON EARLY MARS.

L. L. Brown and J. F. Kasting, Department of Geosciences, Pennsylvania State University, University Park PA 16802, USA.

One explanation for the formation of fluvial surface features on early Mars is that the global average surface temperature was maintained at or above the freezing point of water by the greenhouse warming of a dense CO_2 atmosphere [1]; however, Kasting [2] has shown that CO_2 alone is insufficient because the formation of CO_2 clouds reduces the magnitude of the greenhouse effect. It is possible that other gases, such as NH_3 and CH_4, were present in the early atmosphere of Mars and contributed to the greenhouse effect. Kasting et al. [4] investigated the effect of NH_3 in a CO_2 atmosphere and calculated that an NH_3 mixing ratio of $\sim 5 \times 10^{-4}$ by volume, combined with a CO_2 partial pressure of 4–5 bar, could generate a global average surface temperature of 273 K near 3.8 b.y. ago when the fluvial features are believed to have formed. Atmospheric NH_3 is photochemically converted to N_2 by ultraviolet radiation at wavelengths shortward of 230 nm; maintenance of sufficient NH_3 concentrations would therefore require a source of NH_3 to balance the photolytic destruction. We have used a one-dimensional photochemical model to estimate the magnitude of the NH_3 source required to maintain a given NH_3 concentration in a dense CO_2 atmosphere [5]. We calculate that an NH_3 mixing ratio of 10^{-4} requires a flux of NH_3 on the order of 10^{12} molecules cm^{-2} s^{-1}. This figure is several orders of magnitude greater than estimates of the NH_3 flux on early Mars; thus it appears that NH_3 mixed with CO_2 is not enough to keep early Mars warm.

We are currently using a one-dimensional radiative-convective climate model to determine the greenhouse effect of CH_4 in a CO_2 atmosphere. Atmospheric CH_4 would have a longer lifetime than NH_3 because CH_4 photolysis occurs only at wavelengths shortward of 145 nm, whereas NH_3 is photolyzed out to 230 nm. Hydrocarbon aerosols, which are formed as a product of CH_4 photolysis, are highly absorbent and may have provided a UV shield that would have lengthened the photochemical lifetime of CH_4 itself, as well as other hydrocarbon gases and NH_3. The greenhouse effect resulting from the combination of these gases and particles could conceivably have raised the mean global surface temperature of Mars to near the H_2O freezing point. A combination of radiative-convective climate modeling and photochemical modeling should show whether this idea is feasible and how large a CH_4 source would be needed.

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OXIDATION OF DISSOLVED IRON UNDER WARMER, WETTER CONDITIONS ON MARS: TRANSITIONS TO PRESENT-DAY ARID ENVIRONMENTS.

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Introduction: The copious deposits of ferric-iron assemblages littering the surface of bright regions of Mars indicate that efficient oxidative weathering reactions have taken place during the evolution of the planet. Because the kinetics of atmosphere-surface (gas-solid) reactions are considerably slower than chemical weathering reactions involving an aqueous medium, most of the oxidation products now present in the martian regolith probably formed when groundwater flowed near the surface. This paper examines how chemical weathering reactions were affected by climatic variations when warm, wet environments became arid on Mars. Analogies are drawn with hydrogeochemical and weathering environments on the Australian continent where present-day oxidation of iron is occurring in acidic groundwater under arid conditions.

Background: Chemical weathering reactions of basaltic rocks are facilitated in aqueous solutions. Several stages are involved [1–3], including (1) dissolution of basaltic glass, iron sulfides, and ferromagnesian silicate minerals, which deliver soluble Mg^{2+}, Fe^{2+}, Ca^{2+}, silica, etc., to groundwater; (2) ferrolysis, during which oxidation of dissolved Fe^{2+} occurs, producing soluble Fe^{3+} ions, which are eventually hydrolyzed to Fe(III) oxyhydroxy- and hydroxysulfato gels and colloids; and (3) precipitation of poorly crystalline ferric oxides, oxyhydroxides, and sulphate minerals, as well as clay silicate and evaporite minerals, in depositional environments such as the martian regolith.

Rates of chemical weathering of Fe^{2+}-bearing minerals in aqueous environments that are applicable to the martian surface have been estimated from experimental data for basaltic minerals [2,3]. Reaction rates are strongly influenced by acidity or pH [4,5], as well as salinity or ionic strength [6], concentration of dissolved O in aerated groundwater [7,8], and temperature [7,8]. In acidic groundwater (pH <4.5), silicate minerals dissolve rapidly, but rates of oxidation of aqueous Fe^{2+} ions are very slow, particularly in brines at low temperatures.

For example, rates of dissolution of olivine and pyroxenes range from about 1400 ppm Fe m^{-2} yr^{-1} (pH 2 at 25°C) to 2×10^{-2} ppm Fe m^{-2} yr^{-1} pH 6 at 0°C). In acidic ice-cold saline solutions (pH 4.5 at 0°C), dissolution rates are about 1 ppm Fe m^{-2} yr^{-1}. In such melt waters saturated with O in the present-day martian atmosphere ($P_{O_2} - 10^{-5}$ bar), the rate of oxidation of dissolved Fe^{2+} is also about 1 ppm Fe m^{-2} yr^{-1}. Rates of oxidation are much higher in near-neutral pH saline groundwater and brines; thus, for brines with ionic strengths of 1 to 5 molal, rates of oxidation range from 500 to 900 ppm Fe m^{-2} yr^{-1} (pH 6 at 0°C) to about 100 ppm Fe m^{-2} yr^{-1} (pH 6 at –25°C). Such relatively low rates of oxidation of aqueous Fe^{2+} contrast with the very high values for terrestrial river water ($\approx 1.8 \times 10^7$ ppm Fe m^{-2} yr^{-1} for pH 6 at 25°C), and for deep ocean bottom water ($\approx 5 \times 10^6$ ppm Fe m^{-2} yr^{-1} for pH 8.2 at 2°C). On Mars, the rate of oxidation of dissolved Fe^{2+} in aerated near-neutral pH saline solutions would exceed the rate of supply of dissolved Fe, except in very acidic groundwater.

Calculations indicate that the mixing ratio of O in the present-day martian atmosphere is not being regulated by the oxidation of aqueous Fe^{2+} ions. In pH 6 saline waters at 0°C, rates of oxidation of dissolved Fe^{2+} would have to exceed 14,000 ppm Fe m^{-2} yr^{-1} to maintain the P_{O_2} at 10^{-5} bar relative to $P_{CO_2} = 3.4 \times 10^4$ bar. If the P_{CO_2} was 1000× higher than present-day values (i.e., P_{CO_2} = 0.35 bar), a P_{O_2} of 10^{-5} bar could be achieved, supporting the suggestion [e.g., 9] that atmospheric pressure was higher earlier in Mars' history. On the other hand, the comparatively high P_{O_2} of the martian atmosphere would effectively oxidize the high concentrations of Fe^{2+} ions that could be present in very acidic solutions if such groundwaters occurred on Mars.

Generation of Acidic Groundwater: Oxygen in subsurface groundwater is consumed by oxidation of iron sulfides (e.g., $FeS_2 + 7/2\ O_2 + H_2O \rightarrow Fe^{2+} + 2\ SO_4^{2-} + 2\ H^+$) and by oxidation of some of the Fe^{2+} ions (i.e., $Fe^{2+} + 1/4\ O_2 + 3/2\ H_2O \rightarrow FeOOH + 2\ H^+$) released during oxidation of sulfides and from dissolution of ferromagnesian silicates. Such ferrolysis reactions cause groundwater to become acidic, facilitating the dissolution of ferromagnesian silicates. Since Fe^{2+} ions are stabilized and are very slowly oxidized in low pH saline solutions, concentrations of dissolved ferrous iron increase (perhaps to as high as 1000 ppm) and may persist indefinitely in acidic groundwater (now permafrost on Mars). Oxidation to insoluble nanophase ferric oxides, oxyhydroxides, and hydroxysulfate minerals would occur rapidly when melt waters become oxygenated in contact with the atmosphere and when acid-buffering reactions occurred involving the formation of clay silicates. Such clay silicates include authigenic Mg-Fe saponites, which are also precipitated from acidic brines, and residual montmorillonites derived from leached basaltic feldspars. Environments on the present-day martian surface that are capable of precipitating ferric-bearing assemblages are equatorial melt waters and regions where sublimation of permafrost has induced the oxidation of Fe^{2+} ions after they were released in evaporite deposits. Oxidative weathering reactions in arid environments might have been more prevalent earlier in the history of Mars, however, by analogy with unique terrestrial environments in Australia.

Terrestrial Analogs: On Earth, natural acidic groundwater systems are comparatively rare but occur, nevertheless, when water seeps through mined sulfide and coal deposits. Outflows of such acid mine drainage water are invariably associated with ochrous ferric-bearing assemblages. On a much larger scale, oxidative weathering associated with acidic groundwater is occurring across the southern half of the Australian continent [10].

In the southeastern part of Western Australia, deep weathering of basement igneous rocks in the Yilgarn Block comprising Archean komatiitic basalts has yielded saline groundwater systems, drainage of which in a semi-arid climate has resulted in extensive playas [10–15]. The discharging water contains high concentrations of dissolved Al, Si, and Fe^{2+}, oxidation and hydrolysis of which generates very acidic groundwater (pH ≥ 2.8), as a result of the ferrolysis reactions. The acidity prevents the precipitation of aluminosilicate clay minerals. Instead, jarosite $[KFe_3(SO_4)_2(OH)_6]$-alunite $[KAl_3(SO_4)_2(OH)_6]$ assemblages are precipitated in the pH range 2.8–6 from solutions with ionic strengths ranging from 1 M to 5 M, occurring in evaporite deposits with gypsum and halite. Ferrihydrite, instead of jarosite-alunite assemblages, is deposited from groundwater depleted in dissolved K and Al. A similar situation may exist on Mars where groundwater associated with parent iron-rich komatiitic basaltic rocks may also have low Al and K contents.

Two stages were important in the development of the acidic hydrogeochemical and arid weathering environment of southern Australia. Initially, there were periods of laterization/ferric oxide deposition under a warm humid climate during the Tertiary. The laterite profiles are characterized by a surface duricrust of Fe and Al oxides over deep clay silicate zones depleted of alkali and alkaline Earth elements. Subsequently, periods of aridity and semi-aridity have continued to the present.

An explanation for the acidic saline groundwater systems on such a large scale and only in Australia lies in the recent climatic history of the continent. After the break from Antarctica began at 65 Ma, Australia moved into the subtropic region. The climate through the Eocene was humid and warm, much like that proposed on early Mars, and periods of laterization occurred. Laterite profiles became abundant in Western Australia on deeply weathered bedrock depleted in alkalis and alkaline earth elements, but enriched in Fe, Al, and Si. Little chemical weathering is now occurring on the Australian continent today due to arid conditions. However, periodic discharges of anoxic acidic groundwater into the arid enviroment and oxidation of the dissolved ferrous iron continue to generate ferric-bearing mineral assemblages in contact with the oxygenated atmosphere. A similar scenario may apply to the present-day surface of Mars, in which periodic precipitation of hydronium jarosite and ferrihydrite may be precursors to nanophase hematite identified in bright regions of the martian surface.

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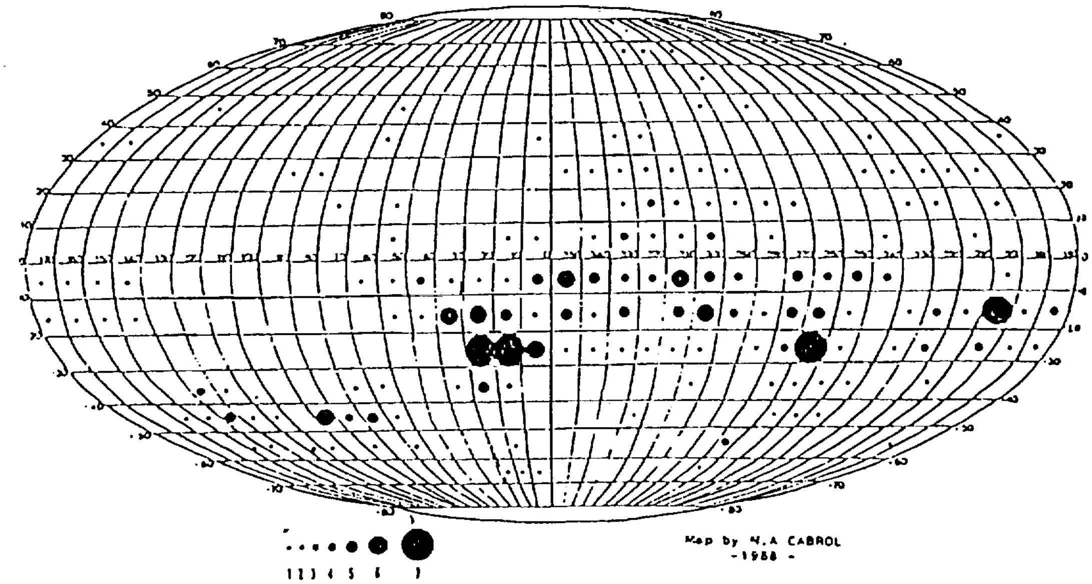
MARS: NOACHIAN HYDROLOGY BY ITS STATISTICS AND TOPOLOGY. N. A. Cabrol and E. A. Grin, Laboratoire de Physique du Système Solaire, Observatoire de Paris-Meudon, 92190, France.

Discrimination between fluvial features generated by surface drainage and subsurface aquifer discharges will provide clues to the understanding of early Mars' climatic history. Our approach is to define the process of formation of the oldest fluvial valleys by statistical and topological analyses [1–3].

Formation of fluvial valley systems reached its highest statistical concentration during the Noachian Period. Nevertheless, they are a scarce phenomenom in martian history, localized on the craterized upland and subject to latitudinal distribution [1,4,5]. They occur sparsely on Noachian geological units with a weak distribution



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1	1–5	5	31–40
2	6–10	6	41–80
3	11–20	7	above 81
4	21–30		

Fig. 1. Distribution of headwater density (hw) by 36×10^4 km².

density (Fig. 1), and appear in reduced isolated surface (around 5×10^3 km²), filled by short streams (100–300 km length) [3].

Topological analysis of the internal organization of 71 surveyed Noachian fluvial valley networks also provides information on the mechanisms of formation.

The tributary hierarchization in these networks shows the specificity of an early martian hydrology. Compared with Earth-like surface drainage systems, which are characterized by a hierarchization of tributaries, martian networks are strongly characterized by typical short theater-headed tributaries with poorly dissected interfluves [6]. The relationship between the number of these headwater branches and the total number of branches of a given network (Fig. 2) provides arguments in favor of the formation of networks highly subject to their terminations. The topological analysis of the internal branch distribution shows common characteristics between most valley networks: (1) over 60% of headwater branches are short tributaries and (2) the relationship between the total branch lengths of a network and its delineation around the headwater system, including its outlet (designated here as drainage density K_d), combined with the planimetric relationship between the circumscribed surface and its perimeter (designated as drainage compacity K_c), shows a good concentration of the physiographic ratio K_d/K_c (Fig. 3). Symbolic of a martian network, the physiographic ratio is a geometric parameter independent of subsurface network geology and topography.

Statistics of terrestrial drainage basins show a mean value $K_d = 0.05$ and $K_c = 1.3$ ($K_d/K_c = 0.04$) compared with the martian value $K_d = 0.1$ and $K_c = 1.4$ ($K_d/K_c = 0.07$). This means that for an equivalent coefficient of compacity K_c, terrestrial drainage systems are twice as dense as the martian ones. The martian physiographic

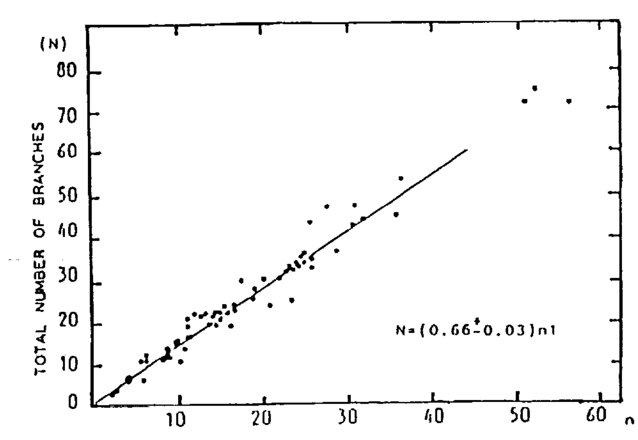

Fig. 2. Relationship between the number of headwater branches (n_1) and total number of branches.

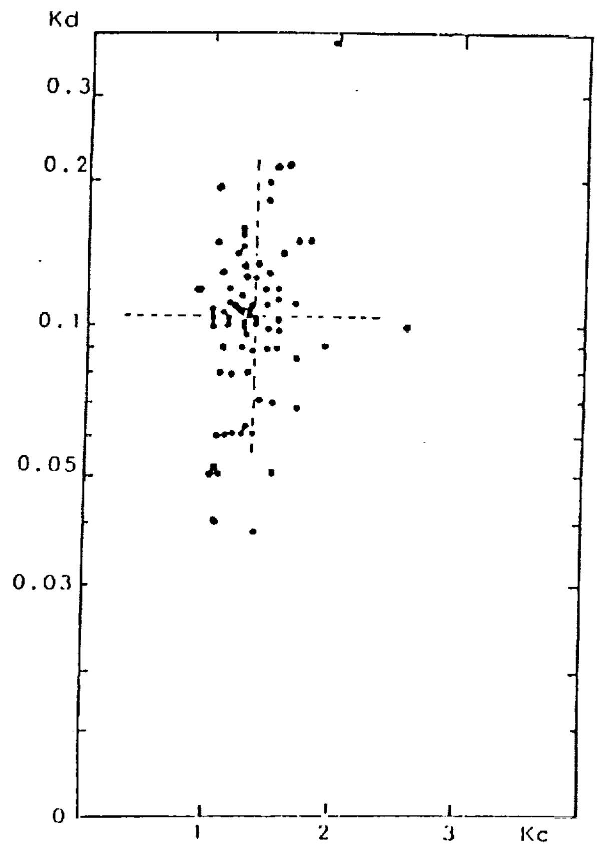

Fig 3. Each network is symbolized by its two physiographic parameters, drainage density (K_d) and drainage basin compacity (K_c).

coefficient is similar to that of terrestrial karst terrain drainage basins. This suggests that martian valley networks are developed on a substratum subject to structural constraints that determine the effectiveness of groundwater flow [7] circulation. The physiographic parameter is not representative of a wide runoff surface but more of reduced-area systems such as sapping valleys, the morphology of which is illustrated by the high frequencies of headwater localized in confined units. Given that major concerns are the history of water and the possibility of life, further investigations of ancient valley networks as unequivocal evidence of circulation of subsurface water will promise a significant advance in our knowledge of Mars.

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THE EVOLUTION OF THE EARLY MARTIAN CLIMATE AND THE INITIAL EMPLACEMENT OF CRUSTAL H_2O.
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Introduction: Given the geomorphic evidence for the widespread occurrence of water and ice in the early martian crust, and the difficulty involved in accounting for this distribution given the present climate, it has been suggested that the planet's early climate was originally more Earth-like, permitting the global emplacement of crustal H_2O by direct precipitation as snow or rain [1,2]. The resemblance of the martian valley networks to terrestrial runoff channels and their almost exclusive occurrence in the planet's ancient (~4-b.y.-old) heavily cratered terrain are often cited as evidence of just such a period. An alternative school of thought suggests that the early climate did not differ substantially from that of today. Advocates of this view find no compelling reason to invoke a warmer, wetter period to explain the origin of the valley networks. Rather, they cite evidence that the primary mechanism of valley formation was groundwater sapping, a process that does not require that surface water exist in equilibrium with the atmosphere [3–5]. However, while sapping may successfully explain the origin of the small valleys, it fails to address how the crust was initially charged with ice as the climate evolved toward its present state. Therefore, given the uncertainty regarding the environmental conditions that prevailed on early Mars, the initial emplacement of ground ice is considered here from two perspectives: (1) The early climate started warm and wet, but gradually cooled with time, and (2) the early climate never differed substantially from that of today.

Early Climate: Warm and Wet: The density and distribution of the valley networks throughout the heavily cratered terrain suggests that, regardless of whether early Mars started warm or cold, groundwater was abundant in the planet's early crust. However, given an initially warm start, an inevitable consequence of both the decline in Mars' internal heat flow and the transition to colder temperatures would have been the development of a freezing front within the regolith that propagated downward with time, creating a thermodynamic sink for any H_2O within the crust. Initially, water may have entered this developing region of frozen ground from both the atmosphere and underlying groundwater. However, as ice condensed within the near-surface pores, it effectively sealed off the deeper regolith from any further atmospheric supply. From that point on, the only source of water for the thickening cryosphere would have been the geothermally driven flux of vapor arising from the presence of groundwater at depth. Indeed, calculations by Clifford [6] indicate that a geothermal gradient as small as 15 K km^{-1} could supply the equivalent of 1 km of water from higher-temperature (higher vapor pressure) depths to the colder (lower vapor pressure) base of the cryosphere every 10^6–10^7 yr. Given the higher geothermal heat flow expected to have characterized the planet 4 b.y. ago, this supply of vapor may have been as much as 3–5 times greater in the past.

Pollack et al. [2] estimate that if the primary mechanism driving climate change was the removal of a massive (1–5-bar) CO_2 atmosphere by carbonate formation, then the transition from a warm to cold early climate must have taken between 1.5×10^7 to 6×10^7 years. For transition times this slow, the downward propagation of the freezing front at the base of the cryosphere is sufficiently small (when compared with the geothermally induced vapor flux arising from the groundwater table) that the cryosphere should have remained saturated with ice throughout its development.

From a mass balance perspective, the thermal evolution of the early crust effectively divided the subsurface inventory of water into two evolving reservoirs: (1) a slowly thickening zone of near-surface ground ice and (2) a deeper region of subpermafrost groundwater. One possible consequence of this evolution is that, if the planet's initial inventory of outgassed water was small, the cryosphere may have eventually grown to the point where all the available H_2O was taken up as ground ice [7]. Alternatively, if the inventory of H_2O exceeds the current pore volume of the cryosphere, then Mars has always had extensive bodies of subpermafrost groundwater. As argued by Clifford [8], this latter possibility is strongly supported by the apparent occurrence of outflow channels as recently as the Mid to Late Amazonian [e.g., 9,10].

Early Climate: Like the Present: Of course, if early Mars was cold from the start, the initial emplacement of ground ice would have differed significantly from that described by the warm scenario. This possibility was first considered by Soderblom and Wenner [7], who suggest that the initial emplacement of crustal H_2O was the result of the direct injection and migration of juvenile water derived from the planet's interior. There are at least two ways in which this emplacement may have occurred. First, by the process of thermal vapor diffusion [6], water exsolved from cooling magmas will migrate from the warmer to colder regions of the crust. Upon reaching the cryosphere, this H_2O will then be distributed throughout the frozen crust by a variety of thermal processes [11]. As a result, any part of the cryosphere that overlies or surrounds an area of magmatic activity will quickly become saturated with ice. The introduction of any additional water will then result in its accumulation as a liquid beneath the frozen crust, where, under the influence of the growing local hydraulic head, it will spread laterally in an effort to reach hydrostatic equilibrium. As this flow expands beneath areas where the cryosphere is not yet fully charged with ice, thermal vapor diffusion [6] and the other thermal processes discussed by Clifford [11] will redistribute H_2O into the frozen crust until its pore volume is either saturated or the local source of groundwater is finally depleted.

However, the fate of water released to the cold martian atmosphere is significantly different. The direct injection of a large quantity of vapor into the atmosphere (e.g., by volcanism) will lead to its condensation as ice on, or within, the surrounding near-surface regolith. As the available pore space in the upper few meters of the regolith is saturated with ice, it will effectively seal off any deeper region of the crust as an area of potential storage. From that point on, any excess vapor that is introduced into the atmosphere will be restricted to condensation and insolation-driven redistribution on the surface until it is eventually cold-trapped at the poles. Should such polar deposition continue, it will ultimately lead to basal melting [12], recycling water back into the crust beneath the caps. As the meltwater accumulates beneath the polar cryosphere, it will create a gradient in hydraulic head that will drive the flow of groundwater away from the poles. As the flow expands radially outward, it will pass beneath regions where, as a result of vapor condensation from the atmosphere, only the top few meters of the cryosphere have been saturated with ice. As before, the presence of a geothermal gradient will then lead to the vertical redistribution of H_2O from the underlying groundwater until the pore volume of the cryosphere is saturated throughout. In this way, the early martian crust may have been globally charged with water and ice without the need to invoke an early period of atmospheric precipitation.

This analysis suggests that, whether the early martian climate started warm or cold, thermal processes within the crust played a critical role in the initial emplacement of ground ice. An important consequence of this fact is that, below the depths of equatorial desiccation predicted by Clifford and Hillel [13] and Fanale et al. [14], the cryosphere has probably been at or near saturation throughout its development, or at least until such time as the total pore volume of the cryosphere grew to exceed the total volume of the planet's outgassed inventory of water. The existence of outflow channels with apparent ages of less than 1 b.y. [9,10] raises considerable doubt as to whether this last stage in the evolution of the martian cryosphere has yet been reached.

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57-91 N94-21666 97 p. 3

THE HYDROLOGIC RESPONSE OF MARS TO THE ONSET OF A COLDER CLIMATE AND TO THE THERMAL EVOLUTION OF ITS EARLY CRUST. S. M. Clifford, Lunar and Planetary Institute, Houston TX 77058, USA.

Morphologic similarities between the martian valley networks and terrestrial runoff channels have been cited as evidence that the early martian climate was originally more Earth-like, with temperatures and pressures high enough to permit the precipitation of H_2O as snow or rain [1,2]. Although unambiguous evidence that Mars once possessed a warmer, wetter climate is lacking, a study of the transition from such conditions to the present climate can benefit our understanding of both the early development of the cryosphere and the various ways in which the current subsurface hydrology of Mars is likely to differ from that of the Earth. Viewed from this perspective, the early hydrologic evolution of Mars is essentially identical to considering the hydrologic response of the Earth to the onset of a global subfreezing climate.

If the valley networks did result from an early period of atmospheric precipitation, then Mars must have once possessed near-surface groundwater flow systems similar to those currently found on Earth, where, as a consequence of atmospheric recharge, the water table conformed to the shape of the local terrain. However, with both the transition to a colder climate and the decline in Mars' internal heat flow, a freezing front eventually developed in the regolith that propagated downward with time, creating a thermodynamic sink for any H_2O within the crust. Initially, water may have entered this developing region of frozen ground from both the atmosphere and underlying groundwater. However, as ice condensed within the near-surface pores, the deeper regolith was ultimately sealed off from any further atmospheric supply. From that point on, the only source of water for the thickening cryosphere must have been thermally driven upward flux of vapor from the underlying groundwater [3,4].

With the elimination of atmospheric recharge, the elevated water tables that once followed the local topography eventually decayed. The continuity of pore space provided by sediments, breccia, and interbasin faults and fractures should have then allowed the water table to hydrostatically readjust on a global scale until it ultimately conformed to a surface of constant geopotential. This conclusion is supported by investigations of areally extensive groundwater systems on Earth that experience little or no precipitation [e.g., 5,6].

The time required for the development of the cryosphere can be calculated by solving the transient one-dimensional heat conduction equation for the case of a semi-infinite half-space with internal heat generation, where

$$\frac{\partial^2 T}{\partial z^2} + \frac{S}{k} + = \frac{1}{a}\frac{\partial T}{\partial t} \qquad (1)$$

and where T is the crustal temperature, z is the depth, t is time, k is the crustal thermal conductivity, a is the thermal diffusivity ($= k/\rho c$), and S is the heat generation rate per unit volume ($= 3Q_g/R$, where Q_g is the geothermal heat flux and R is the radius of Mars) [7]. An upper limit can be placed on how rapidly the cryosphere evolved if we assume that the surface temperature of Mars underwent an instantaneous transition from a mean global value of 273 K to its current latitudinal range of 154–218 K. For these conditions, equation (1) was solved numerically using the method of finite differences. The results indicate that, given a present-day geothermal heat flux of 30 mW m^{-2}, the freezing front at the base of the cryosphere will reach its equilibrium depth (~3700 m) at the equator in $\sim 4.6 \times 10^5$ yr, while at the poles it will take roughly 1.5×10^6 yr ($z = 7900$ m). Given the elevated geothermal conditions that probably characterized the planet 4 b.y. ago (i.e., $Q_g \sim 150$ mW m^{-2}), the corresponding development times are 2×10^4 yr ($z = 700$ m) and 1.3×10^5 yr ($z = 1600$ m) respectively. These calculations are based on a thermal conductivity of 2.0 W m^{-1} K^{-1}, a freezing temperature of 273 K, and a maximum latent heat release (due to the condensation of H_2O vapor as ice in the pores) that does not exceed Q_g, a limit imposed by the geothermal origin of the vapor flux reaching the base of the cryosphere [e.g., 3,4]. Note that although these development times assume an initially dry crust, they would not be significantly different even if the crust were initially saturated throughout. Although the early growth of the cryosphere would be slowed by the ready supply of latent heat, this period represents only a small fraction of the total time required for the cryosphere to reach equilibrium. In the later stages of growth, which are controlled almost exclusively by conduction through the frozen crust, the rate of heat loss is sufficiently small that the effect of latent heat release can be virtually ignored.

Although the assumption that Mars underwent an instantaneous transition from a warm to cold early climate is clearly incorrect, this extreme example serves to illustrate an important point, i.e., on a timescale greater than $\sim 10^6$ yr, the base of the cryosphere is essentially in thermal equilibrium with mean temperature environment at the surface. As a result, for any reasonable model of climate evolution, the growth of the cryosphere is not controlled by the rate of conduction through the crust, but by how rapidly the mean surface temperature environment changes with time. Pollack et al. [2] estimate that if the primary mechanism driving climate change was the removal of a massive (1–5-bar) CO_2 atmosphere by carbonate formation, then the transition from a warm to cold early climate must have taken between 1.5×10^7 and 6×10^7 yr. For transition times this slow, the downward propagation of the freezing front at the base of the cryosphere proceeds at a rate that is sufficiently small (when compared with the geothermally induced vapor flux arising from the groundwater table) that the geothermal gradient should have no trouble supplying enough vapor to keep the cryosphere saturated with ice throughout its development.

From a mass balance perspective, the thermal evolution of the early crust effectively divided the subsurface inventory of water into two reservoirs: (1) a slowly thickening zone of near-surface ground ice and (2) a deeper region of subpermafrost groundwater [8]. Regardless of how rapid the transition to a colder climate actually was, the cryosphere has continued to thicken as the geothermal output from the planet's interior has gradually declined. One possible consequence of this evolution is that, if the planet's initial inventory of outgassed water was small, the cryosphere may have eventually grown to the point where all the available H_2O was taken up as ground ice [9]. Alternatively, if the inventory of H_2O exceeds the current pore volume of the cryosphere, then Mars has always had extensive bodies of subpermafrost groundwater.

Because the pore volume of the cryosphere was probably saturated with ice throughout its early development, the thermally driven vapor flux arising from the reservoir of underlying groundwater could have led to the formation and maintenance of near-surface perched aquifers, fed by the downward percolation of condensed vapor from the higher and cooler regions of the crust. Eventually the hydrostatic pressure exerted by the accumulated water may have been sufficient to disrupt the overlying ground ice, allowing the stored volume to discharge onto the surface. Such a scenario may have been repeated hundreds of times during the planet's first 500 m.y. of geologic history, possibly explaining (in combination with local hydrothermal systems driven by impact melt [10,11] and volcanism [12]) how some valley networks may have evolved in the absence of atmospheric precipitation [13,14]. However, as the internal heat flow of the planet continued to decline, the thickness of the cryosphere may have grown to the point where it could no longer be disrupted by the limited hydrostatic pressure that could develop in a perched aquifer, thus terminating the potential contribution of low-temperature hydrothermal convection to valley network formation.

Finally, the postcryosphere groundwater hydrology of Mars will differ from its possible precryosphere predecessor (and therefore from present-day terrestrial groundwater systems) in at least one other important way. In contrast to the local dynamic cycling of near-surface groundwater that may have characterized the first 500 m.y. of martian climate history, the postcryosphere period will necessarily be dominated by deeper, slower interbasin flow. Aside from polar basal melting, there are at least three other processes that are likely to drive flow under these conditions: (1) tectonic uplift (essentially the same mechanism proposed by Carr [8] to explain the origin of the outflow channels east of Tharsis), (2) gravitational compaction of aquifer pore space (perhaps aided by the accumulation of thick layers of sediment and basalt on the surface), and (3) regional-scale hydrothermal convection (e.g., associated with major volcanic centers such as Tharsis and Elysium). Note that, with the exception of active geothermal areas, the flow velocities associated with these processes are likely to be orders of magnitude smaller than those that characterize precipitation-driven systems on Earth.

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THE EARLY MARTIAN ENVIRONMENT: CLUES FROM THE CRATERED HIGHLANDS AND THE PRECAMBRIAN EARTH.

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There is abundant geomorphic evidence to suggest that Mars once had a much denser and warmer atmosphere than present today. Outflow channels [1], ancient valley networks [2], and degraded impact craters in the highlands [3] all suggest that ancient martian atmospheric conditions supported liquid water on the surface. The pressure, composition, and duration of this atmosphere is largely unknown. However, we have attempted to place some constraints on the nature of the early martian atmosphere by analyzing morphologic variations of highland impact crater populations, synthesizing results of other investigators, and incorporating what is known about the geologic history of the early Earth. This is important for understanding the climatic evolution of Mars, the relative abundance of martian volatiles, and the nature of highland surface materials.

The duration of the martian primordial atmosphere and the interval of time water existed as a liquid on the surface can be estimated from the ages of features thought to have formed by fluvial processes. Formation of the large outflow channels occurred from the late Noachian (e.g., Ma'adim Vallis [4]) until the early Amazonian (e.g., Mangala Valles [5]). Formation of the ancient valley networks [6] and degradation of the cratered highlands [3] also occurred during this period, and the timing of both these processes appears to be dependent upon elevation. Formation of valley networks and degradation of highland impact craters ceased at higher elevations before they shut off at lower elevations. This implies that the highland volatile reservoir became depleted with time or, alternatively, the density of the martian atmosphere decreased with time. In the latter scenario, precipitation would occur throughout the highlands initially fairly independent of elevation. With time and loss of atmosphere, cloud condensation (and thus precipitation) could occur only at progressively lower altitudes. Condensation of the Earth's primordial steam atmosphere occurred ~4.0 b.y. ago [7]. Assuming that the martian primordial atmosphere also condensed at approximately the same time (perhaps sooner given Mars' distance from the Sun), the maximum interval of time liquid water existed on the surface is either ~1.2 b.y. or ~450 m.y. The differences in these estimates are due to the uncertainties in the absolute ages of the martian periods and are based on two different models of the cratering flux at Mars [8,9 respectively].

Because the early Sun is thought to have had a lower luminosity than today [10], ~5 bar of CO_2 may have been needed to maintain the ancient martian surface temperature above freezing [11]. On the other hand, models incorporating early solar mass loss [12] suggest that early solar luminosities were actually much higher than today. This would allow the early martian atmosphere to be much thinner (~1 bar) and yet warm enough for liquid water. Regardless, the age and elevation relationship of features contained in the cratered highlands suggest that the pressure of the martian primordial atmosphere was never fixed at a high level but steadily decreased to <1 bar at about the beginning of the Amazonian. If sapping and seepage of groundwater were the mechanisms for ancient valley network formation [2] and highland degradation [3], a recharge mechanism is still needed for the aquifer in order to maintain these processes over the interval of time they were operating (tens to hundreds of millions of years at any given elevation). Again, martian atmospheric pressures must have been maintained >1 bar for a long time in order to produce the weather patterns necessary to efficiently recharge the highland aquifer at higher elevations. Based on the amount of time fluvial processes occurred on Mars, a rough estimate of the primordial atmospheric pressure is ~5–10 bar. Possible recharge of the atmosphere through impact-induced CO_2 [13] suggests that highland degradation may have also been periodic, but such a mechanism would have become less efficient with time to correlate with the age/elevation relations of highland features. If the weak Sun paradox is wrong [11], the primordial atmosphere would still need to be between ~5 and 10 bar initially to allow ancient valley network formation and highland degradation to occur over the interval of time observed.

The primordial terrestrial atmosphere and oceans were highly reducing [14]. Similarly, it is probable that Mars also had a primordial, highly reducing atmosphere and surface waters. Possible release of martian water in a CO_2-rich atmosphere by precipitation and channel-forming processes has led numerous investigators to speculate on the creation of massive martian carbonate deposits [e.g., 11]. The formation of such deposits would tend to remove CO_2 from the martian atmosphere and would require a substantially thick primordial atmosphere (~20 bar [11]) for ancient valley network and highland degradation processes to operate for ~450 m.y. or ~1.2 b.y. However, these hypotheses have been based on the assumption that an acidic primordial atmosphere (pH <1–3 [15]) was buffered by cations released through the weathering of rock. Precipitation of calcium carbonate occurs only in water with a high pH (>7.8 [16]). On Earth, precipitation of calcium carbonate did not become a common phenomenon until the Proterozoic (2.5 b.y.) when stable, shallow marine cratons developed, thus allowing weathered cations to concentrate [17]. Simply, terrestrial oceans were in existence ~1.5 b.y. before precipitation of calcium carbonate started to occur, and even then it was in unique circumstances!

On Mars two scenarios exist for standing bodies of water, the extreme being Oceanus Borealis [18]. This large ocean, thought to occupy much of the northern plains, postdates most of the fluvial features on Mars (up to the middle Amazonian), is needed to explain

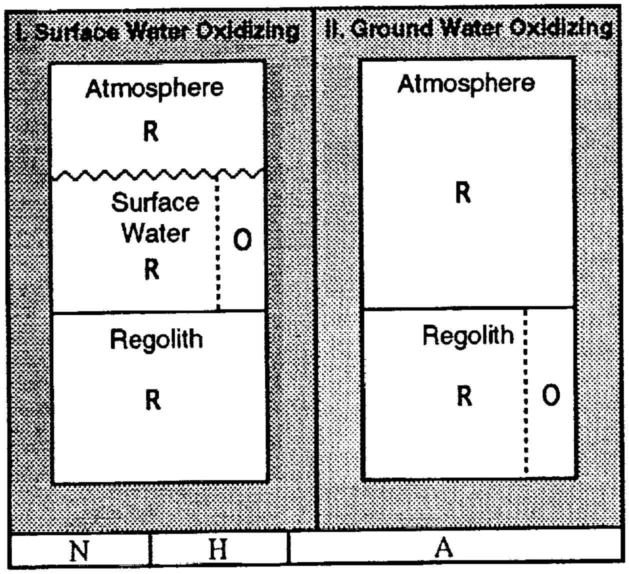

Fig. 1. A two-box model showing the location of reducing (R) and oxidizing (O) conditions on Mars through time. Oxidizing conditions in stage I represent local O oases, which may have developed in small pools formed by precipitation or seepage. Oxidizing conditions in stage II represent buried O oases located where groundwater is enriched by volcanics from an evolved mantle. Based on Fig. 3 in [14].

the formation of putative glacial features in the southern highlands, and probably did not exist (see discussion in [19]). More realistic are the smaller, isolated lacustrine basins proposed for low-lying areas in the northern plains surrounded by outflow channels [20,21] and in the cratered highlands where numerous ancient valley networks terminate [22]. Such basins are supported by geomorphic evidence (e.g., possible wave-cut terraces [20,21]) and occur during the interval that other fluvial features were forming. In addition to these basins, smaller isolated pools may have developed in topographic depressions (e.g., impact craters) if precipitation occurred in the highlands as suggested in [3]. It is especially likely that these smaller pools allowed eroded materials to concentrate, raise the local pH, and induce the formation of carbonates. Such "oases" could be the most likely locations of carbonate deposits on Mars. Other oases may exist underground where oxidizing volcanic gases interact with groundwater to raise the pH, and may help explain the presence of carbonates in the SNC meteorites [23]. Based on this view and the terrestrial example [14], a "two-box" model for the oxidizing stages for Mars is proposed (Fig. 1).

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AN ATTEMPT TO COMPREHEND MARTIAN WEATHERING CONDITIONS THROUGH THE ANALYSIS OF TERRESTRIAL PALAGONITE SAMPLES. C. Douglas[1], I. P. Wright[1], J. B. Bell[2], R. V. Morris[3], D. C. Golden[3], and C. T. Pillinger[1], [1]The Open University, Walton Hall, Milton Keynes, MK7 6AA, UK, [2]Mail Stop 245-3, NASA Ames Research Center, Moffet Field CA 94035-1000, USA, [3]NASA Johnson Space Center, Houston TX 77058, USA.

Spectroscopic observations of the martian surface in the visible to near infrared (0.4–1.0 μm), coupled with measurements made by Viking, have shown that the surface is composed of a mixture of fine-grained weathered and nonweathered minerals. The majority of the weathered components are thought to be materials like smectite clays, scapolite, or palagonite [1]. Until materials are returned for analysis there are two possible ways of proceeding with an investigation of martian surface processes: (1) the study of weathering products in meteorites that have a martian origin (SNCs) and (2) the analysis of certain terrestrial weathering products as analogs to the material found in SNCs, or predicted to be present on the martian surface. Herein, we describe some preliminary measurements of the carbon chemistry of terrestrial palagonite samples that exhibit spectroscopic similarities with the martian surface [2–4]. The data should aid the understanding of weathering in SNCs and comparisons between terrestrial palagonites and the martian surface.

The SNC meteorites contain a variety of weathering products including carbonates, sulphates, and clays of martian and terrestrial origin [5–7]. The C chemistry of SNC meteorites has already been studied extensively so it would seem reasonable to characterize the nature of C in terrestrial palagonites with a view to comparing the data with the results from SNC meteorites [6]. Several mechanisms for the formation of palagonites (which are hydrated and devitrified basaltic glasses) have been proposed: (1) hydrothermal alteration (induced by volcanism, geothermal gradients, or impact), (2) subpermafrost magmatic intrusion, (3) subaerial intrusion above the permafrost layer, and (4) static gas-solid weathering [8].

As a preliminary investigation, two terrestrial palagonite samples have been analyzed for C stable isotopes by the use of stepped combustion and static mass spectrometry. One, PN-9, collected from Mauna Kea, is a palagonitic soil and has been studied extensively because its infrared absorption features closely resemble that of the martian surface [2,5,9]. The C released between room temperature

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and 1200°C accounted for 2.2 wt% of the sample, far in excess of even the most C-rich SNCs; clearly care must be taken when using palagonites as analogs of martian weathering products. The isotopic release profile of PN-9 shows at least three distinct C components. The first is a heavy component, reaching a $\delta^{13}C$ of –9.5‰, and released between room temperature and 200°C, accounting for 210 ± 10 ppm. This low-temperature C may be a loosely bound labile material or it may be adsorbed atmospheric CO_2. Atmospheric CO_2 will be adsorbed more easily onto fine-grained material and so analysis of size separates may prove interesting. A second component is released between 300°C and 450°C, which reaches a $\delta^{13}C$ maximum of –21.4‰; this represents a lower limit to the actual isotopic composition of this component because of a concomitant release of third-component organic materials ($\delta^{13}C$ = –25‰) over the temperature range 200°–600°C. Carbon components with a similar release temperature and isotopic composition have been observed in SNCs on previous occasions; e.g., LEW 88516, (subsamples ,8 and ,13) have distinct C components of $\delta^{13}C$ of –19.5‰ and –21.6‰ respectively across the 250°–450°C temperature interval [10]. This possibly coincidental similarity between SNCs and palagonites deserves further study.

The other palagonite sample analyzed (PH-1) was collected from the Puu Huluhulu cinder cone on Mauna Kea, immediately adjacent to an intruded lava slab [8]. It consists of a thermally altered palagonite tephra containing small amounts of hematite formed from the recrystallization of Fe material during the emplacement of the slab. Hematite is thought to be present on the surface of Mars in small quantities and so PH-1, which contains crystalline hematite, is in better agreement with the martian spectral observations [8]. Analysis of PH-1 showed it to contain 4× less C (0.6 wt%) than PN-9. The lower C content may result from volatile loss during emplacement of the lava slab, or alternatively the high C content of PN-9 may result from the input of organics during soil formation. Analysis of PH-1 showed the presence of a C component below 300°C with a $\delta^{13}C$ of at least –23.5‰. It is possible that this may be a mixture of components: the remains of a small amount of the low-temperature heavy component released in PN-9 and organics of lower $\delta^{13}C$.

At present the study of the low-temperature C components in the palagonites is at an early stage. However, further analyses together with studies of smectite and montmorillinite clays may help to clarify the situation. A complete understanding of the low-temperature weathering products produced on Earth will ultimately help constrain the operation of atmospheric and liquid phase reactions occurring on Mars. A major problem for the interpretation of data from SNC meteorites is in distinguishing terrestrial and preterrestrial weathering products. It is hoped that this particular study will help solve some of the difficulties.

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REQUIREMENTS FOR THE EARLY ATMOSPHERE OF MARS FROM NITROGEN ISOTOPE RATIOS. J. L. Fox, Institute for Terrestrial and Planetary Atmospheres, State University of New York at Stony Brook, Stony Brook NY 11794, USA.

The N escape models of Fox and Dalgarno [1] and Fox [2] required the presence of a dense, early CO_2 atmosphere to inhibit fractionation of the N isotopes ^{15}N and ^{14}N. The computed photochemical escape fluxes are so large at the present that the isotope ratio measured by Viking (about 1.62× terrestrial) can be produced in about 1.5 b.y. We have refined this model in several ways. It has been updated to incorporate the variation of the escape fluxes with increases in the solar fluxes at earlier times according to the model of Zahnle and Walker [3]. As expected, this exacerbates the problem with overfractionation, but not greatly. Most of the escape and fractionation of the N occurs in the last 1.5 b.y., when the solar flux was only slightly different from the present. The dense early atmosphere must persist only a bit longer in order to reproduce the measured isotope ratio. We have also modified the model to take into account changes in the O mixing ratio with time in the past, assuming that the O abundance is proportional to the square root of the solar flux. Although the production rate of O from photodissociation of CO_2 scales as the solar flux, the strength of the winds and other mixing processes also increases with the solar flux [4], resulting in possibly more effective transport of O to the lower atmosphere where it is destroyed by catalytic and three-body recombination mechanisms. The escape fluxes due to the ion-neutral reactions

$$N_2^+ + O \rightarrow NO^+ + N \tag{1}$$

and

$$O^{++} + N_2 \rightarrow N^+ + N + O^+ \tag{2}$$

are thus changed slightly at earlier times compared to models in which the O mixing ratio is assumed to be proportional only to the CO_2 mixing ratio.

The role of dissociative recombination of N_2^+

$$N_2^+ + e \rightarrow N + N \tag{3}$$

is important because it involves an inherent fractionation mechanism in addition to that produced by diffusive separation between the homopause and the exobase. Previously we have assumed that the rate of dissociative recombination at earlier times scales as the mixing ratio of N_2 at the exobase. Although this is a good approximation for small mixing ratios of N_2, it is an overestimate for large mixing ratios. Ion-neutral chemistry tends to transform ions whose parent neutrals have high ionization potentials, such as N_2^+, into ions whose parents have lower ionization potentials. The ratio of exobase densities of N_2^+ to those of other ions are thus smaller than the ratio of the production rates. For example, we have found that if the mixing ratio of N_2 is 0.75 at the homopause, it is about 0.95 at the exobase, but N_2^+ constitutes only half the total ion density at the exobase. Thus the dissociative recombination rate and the fractionation due to dissociative recombination are slightly reduced at earlier times. Another possibility that we will explore for reducing the fractionation in dissociative recombination is to incorporate the higher electron temperatures that have been indicated by a recent analysis of Viking RPA data [5]. Higher electron temperatures provide more energy for the ^{15}N atoms released in dissociative recombination of $^{15}N/^{14}N$ at the exobase, and thus the escaping fraction is larger than that computed by Wallis [6]. Luhmann et al. [7] have computed the sputtering rates of atmospheric O and C by O^+ ions picked up by the solar wind. The addition of sputtering as a loss process for N_2 greatly exacerbates the problem with overfractionation of $^{15}N/^{14}N$ [8]. We find that even a dense, early atmosphere cannot inhibit the enormous escape rates and subsequent fractionation implied by the Luhmann et al. fluxes.

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FLUVIAL VALLEYS IN THE HEAVILY CRATERED TERRAINS OF MARS: EVIDENCE FOR PALEOCLIMATIC CHANGE? V. C. Gulick and V. R. Baker, Department of Geosciences and the Lunar and Planetary Laboratory, University of Arizona, Tucson AZ 85721, USA.

Whether the formation of the martian valley networks provides unequivocal evidence for drastically different climatic conditions remains debatable. Recent theoretical climate modeling precludes the existence of a temperate climate early in Mars' geological history [1]. An alternative hypothesis [2] suggests that Mars had a globally higher heat flow early in its geological history, bringing water tables to within 350 m of the surface. While a globally higher heat flow would initiate groundwater circulation at depth, the valley networks probably required water tables to be even closer to the surface. Additionally, we have previously reported that the clustered distribution of the valley networks within terrain types, particularly in the heavily cratered highlands [3], suggests regional hydrological processes were important. In this abstract, we summarize the case for localized hydrothermal systems and present estimates of both erosion volumes and of the implied water volumes for several martian valley systems.

Sustained groundwater outflow requires that hydraulic gradients be maintained. On Earth, rainfall or melting snow or ice eventually infiltrate into the subsurface and maintain these gradients. Thus on Earth, groundwater outflow and surface runoff are intimately connected and such a connection is reflected in the formation of fluvial systems. In locations where sapping valleys do form they are associated with runoff-dominated systems, regardless of lithologic or climatic conditions [3].

On Mars, however, it is not clear how hydraulic gradients were maintained, particularly in the southern highlands, where most fluvial valleys exhibit a sapping morphology. In these regions, sapping valleys generally do not form together with runoff valleys, but instead form as isolated systems. Thus, groundwater outflow does not seem closely linked to an atmospheric hydrological cycle. In the heavily cratered terrains, evidence for fluvial erosion is found on the ejecta blankets of impact craters, on some volcanos, and in intercrater plains regions. Many valleys in the intercrater plains are associated with dark units that have been interpreted as igneous sill intrusions [4]. An asymmetric distribution of valleys around impact craters is common on Mars, unlike drainages situated around terrestrial impact craters that tend to be more uniformly distributed. While most martian valley networks are attributed to formation by groundwater outflow processes [5–7], the distribution of these networks is unlike that formed by terrestrial sapping valleys.

Lacking an atmospheric hydrologic cycle, subsurface energy sources must maintain hydraulic gradients. Two possibilities are a global, uniformly higher heat flow and localized energy sources, such as magmatic intrusions. Although a global, higher heat flow would produce vertical temperature gradients, it would not produce anomalously large, localized horizontal temperature gradients in the groundwater by itself. Such gradients are necessary to produce lateral flow and recharge of aquifers. However, the addition of vigorous, localized hydrothermal circulation to a uniformly higher heat flow overcomes this problem. Such systems would naturally be associated with igneous intrusions, volcano formation, and large impact craters, all of which are locales for valley formation on Mars, particularly in the heavily cratered terrains. Depending on the volume of the associated magmatic intrusion, martian hydrothermal systems can circulate groundwater into the surface environment for several million years; such systems are thus able to maintain hydraulic gradients sufficient for valley formation. Rather than replenishing groundwater through rainfall and infiltration, our numerical modeling demonstrates that a martian hydrothermal system replenishes itself by continually drawing in colder, denser groundwater radially from more distant parts of the aquifer. The total quantity of groundwater that passes through the modeled hydrothermal system over its lifetime is comparable to that needed to form a single outflow channel. Hence, subsurface aquifers of the required magnitude to form fluvial valleys must have existed on Mars.

The clustered distribution or localization of sapping valleys on Mars and their isolation from runoff valleys strongly suggests localized, subsurface sources of water. In short, a rainfall genesis should produce associated runoff valleys and a more uniform distribution of fluvial valleys within a given terrain type or surface

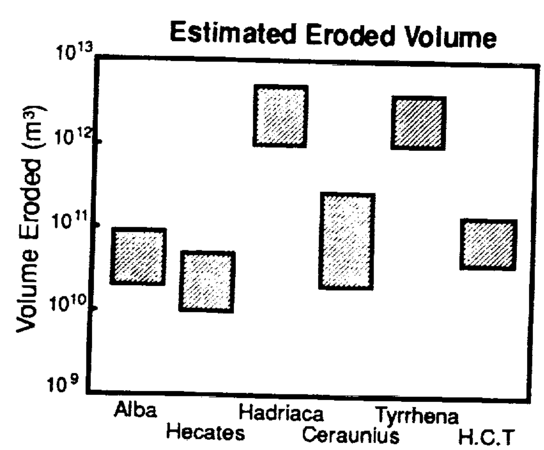

Fig. 1.

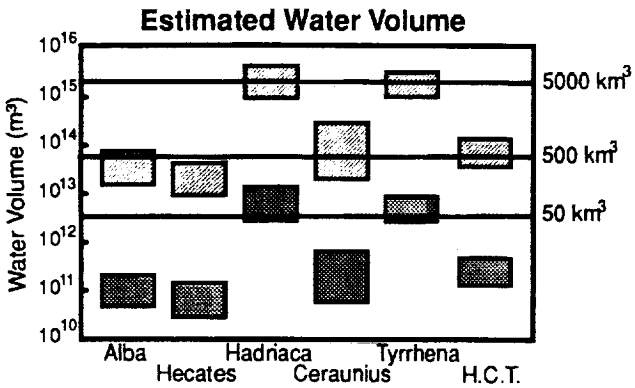

Fig. 2.

geologic unit [3]. A uniform, globally higher heat flow should not produce such localized sapping valleys. It is for these reasons that we invoke localized hydrothermal systems for martian valley genesis. Such hydrothermal systems would be localized in surface extent, yet (as the simulations suggest) draw groundwater from great distances while focusing outflow into relatively small regions. The hydrothermal discharge would also preferentially produce landforms associated with groundwater outflow—the sapping valleys. However, snowfall melting in hydrothermal areas and sublimating elsewhere might also play a role in producing such a distribution [3].

The principle finding of our numerical model is that magmatic intrusions of several 10^2 km^3 provide sufficient volumes of groundwater outflow over the timescales (several 10^5 yr or more) needed to form fluvial valleys [8]. Calculated discharges are robust. Subsurface inhomogeneities, local impermeable caps, and uncertainties in porosity all affect the discharges at the 20% level or less. The parameter with the single greatest effect on the calculated discharges is the subsurface permeability. Permeabilities between 10 and 1000 darcy provide sufficient quantities of groundwater outflow to form fluvial valleys. Lower permeabilities require larger intrusion volumes to produce the same discharge. However, in the range of permeabilities expected for basaltic rock, there is no difficulty in producing significant groundwater outflow.

We have estimated the eroded volumes of two of the best-developed valley networks (Parana and Warrego Valles) in the heavily cratered terrains. These values (H.C.T) are compared with our estimated valley volumes on martian volcanos in Fig. 1. Estimates of martian valley erosion can be combined with terrestrial fluvial erosion rates to obtain estimates of the total volume of water required to form each set of martian valleys. Some ratios of water volume to eroded volume for Mars are as low as 2 or 3 to 1 [9]. However, based upon our own study of fluvial erosion on volcanic landscapes, we find ratios as large as 1000:1. The total water volume using each ratio is shown for each valley group in Fig. 2. For each locality the lower bar represents the uncertainty due to valley side-wall slopes while using a water-to-eroded-volume ratio of 3:1, the upper bar using a ratio of 1000:1. Horizontal lines in Fig. 2 illustrate the cumulative discharge of hydrothermal systems associated with 50-, 500-, and 5000-km^3 igneous intrusions. Therefore we conclude that hydrothermal systems can provide the volumes of groundwater outflow needed to form martian valley networks and can provide an alternative to rainfall from a warm, wet early Mars.

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A MODEL FOR THE EVOLUTION OF CO_2 ON MARS.

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There are several lines of evidence that suggest early Mars was warmer and wetter than it is at present [1]. Perhaps the most convincing of these are the valley networks and degraded craters that characterize much of the ancient terrains. In both cases, fluvial activity associated with liquid water is believed to be involved. Thus, Mars appears to have had a warmer climate early in its history than it does today. How much warmer is not clear, but a common perception has been that global mean surface temperatures must have been near freezing—almost 55 K warmer than at present.

The most plausible way to increase surface temperatures is through the greenhouse effect, and the most plausible greenhouse gas is CO_2. Pollack et al. [2] estimate that in the presence of the faint young Sun, the early martian atmosphere would have to contain almost 5 bar of CO_2 to raise the mean surface temperature up to the freezing level; only 1 bar would be required if the fluvial features were formed near the equator at perihelion at maximum eccentricity. However, these calculations now appear to be wrong since Kasting [3] has shown that CO_2 will condense in the atmosphere at these pressures and that this greatly reduces the greenhouse effect of a pure CO_2 atmosphere. He suggested that alternative greenhouse gases, such as CH_4 or NH_3, are required.

In this paper, we approach the early Mars dilemma from a slightly different point of view. In particular, we have constructed a model for the evolution of CO_2 on Mars that draws upon published processes that affect such evolution. Thus, the model accounts for the variation of solar luminosity with time, the greenhouse effect, regolith uptake, polar cap formation, escape, and weathering. We initialize the model 3.8 G.y. ago with a specified CO_2 inventory and then march it forward in time to the present epoch. The model partitions CO_2 between its various reservoirs (atmosphere, caps, regolith, carbonates, and space) according to the thermal environment predicted by a modified version of the Gierasch and Toon [4] energy balance climate model. The goal is to determine if it is possible to find an evolutionary scenario that is consistent with early fluvial activity, and that arrives at the present epoch with the initial CO_2 partitioned into its various reservoirs in plausible amounts. Our early fluvial activity criterion is that global mean temperatures must be at least 240 K at the beginning of the simulation; our current reservoir criteria is that the atmosphere must hold about 7 mbar of CO_2, the caps several millibars, and the regolith 300 mbar. We do not constrain the final size of the rock reservoir.

We find no evolutionary scenario that satisfies these criterion when using published estimates of the processes involved. The main

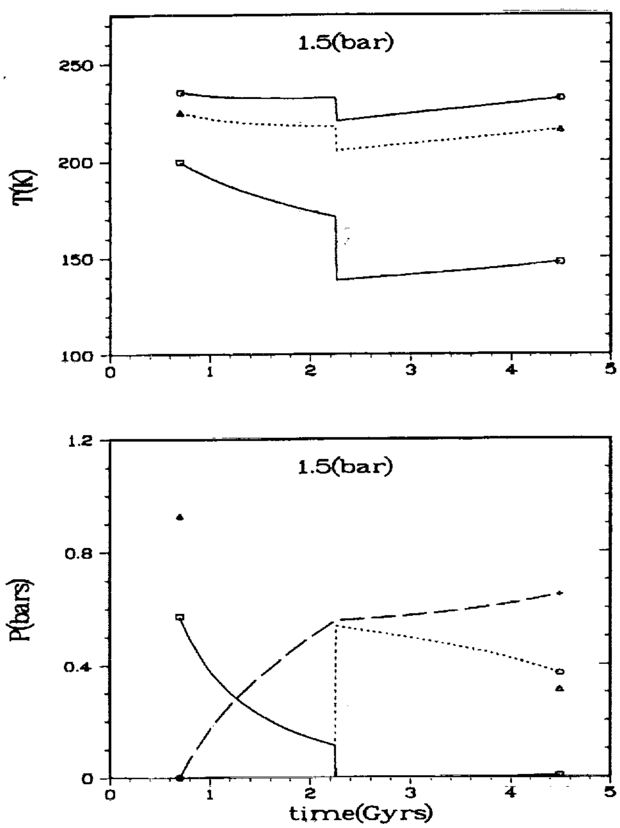

Fig. 1. Temperature (top) and pressure evolution (bottom) for total initial abundance of 1.5 bar. The temperatures shown are for the surface at equator (solid line with circles), pole (solid line with squares), global mean (dashed line with triangles), and frost point (dotted). The CO_2 reservoir pressures are atmosphere (solid line with squares), caps (dashed line with circles), regolith (dotted line with triangles), and carbonate (chain dotted line with plus sign).

difficulty is making Mars warm early on. As Kasting has pointed out, only a stronger greenhouse or a brighter early Sun can help in this regard. However, we have found that if the greenhouse were stronger or if the Sun were brighter, then massive permanent caps would form as a result of a collapse in the climate system sometime between 1.5 and 3.5 b.y. ago. An example of this collapse and the thermal history associated with it is shown in Fig. 1.

The collapse is a result of an unstable feedback between the poleward transport of heat by the atmosphere, the greenhouse effect, and surface pressure. As surface pressure falls, heat transport and the greenhouse effect are reduced, the polar caps cool, surface pressure falls further, and so on. Gierasch and Toon [4] discuss this instability in detail. In our model, the instability is set off by weathering that removes CO_2 from the atmosphere at a rate that is exponentially proportional to temperature. Thus, the warmer early Mars is, the more likely collapse will occur. As much as 600 mbar goes into the caps when collapse occurs with the CO_2 coming from the atmosphere and the regolith. More importantly, at least 300 mbar survives to the present epoch—much more than appears to reside in the south residual cap.

Collapse can be avoided if the polar albedo is significantly lower than the value we have assumed (0.75), or if the actual poleward heat flux is greater than that given by our simple parameterization. However, in either case, the implication is that if global mean surface temperatures were at or above 240 K on early Mars, then a minimum total inventory of 2 bar of CO_2 is required, and at least 70% of it has been sequestered as carbonate in near-surface materials. On the other hand, if the fluvial features in the ancient terrains do not require global mean temperatures near 240 K and can be explained by phenomena that are not climate related, such as an elevated geothermal heat flux, then our model favors an initial CO_2 inventory near 600 mbar. Of this initial CO_2, most has gone into the regolith (300 mbar), modest amounts into carbonates (130 mbar), even smaller amounts into the atmosphere (7 mbar) and caps (3 mbar), with the remainder having escaped into space (160 mbar). Thus, it is crucial that we obtain better constraints on the thermal regime required to form the fluvial features on early Mars.

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MARS ATMOSPHERIC LOSS AND ISOTOPIC FRACTIONATION BY SOLAR-WIND-INDUCED SPUTTERING AND PHOTOCHEMICAL ESCAPE. B. M. Jakosky[1], R. O. Pepin[2], R. E. Johnson[3], and J. L. Fox[4], [1]Laboratory for Atmospheric and Space Physics, University of Colorado, Boulder CO 80309-0392, USA, [2]School of Physics and Astronomy, University of Minnesota, Minneapolis MN 55455, USA, [3]Department of Nuclear Engineering and Engineering Physics, University of Virginia, Charlottesville VA 22903, USA, [4]Department of Mechanical Engineering, State University of New York–Stony Brook, Stony Brook NY 11794, USA.

We examine the effects of loss of Mars atmospheric constituents by solar-wind-induced sputtering and by photochemical escape during the last 3.8 b.y. Sputtering is capable of efficiently removing all species from the upper atmosphere, including the light noble gases; N also is removed by photochemical processes. Due to the diffusive separation by mass above the homopause, removal from the top of the atmosphere will fractionate the isotopes of each species, with the lighter mass being preferentially lost. For C and O, this allows us to determine the size of nonatmospheric reservoirs that mix with the atmosphere; these reservoirs can be accounted for by exchange with CO_2 adsorbed in the regolith and with H_2O in the polar ice deposits. We have constructed both simple analytical models and time-dependent models of the loss of volatiles from and supply to the martian atmosphere. Both Ar and Ne require continued replenishment from outgassing over geologic time.

For Ar, sputtering loss then explains the fractionation of $^{36}Ar/^{38}Ar$ without requiring a distinct epoch of hydrodynamic escape (although fractionation of Xe isotopes still requires a very early hydrodynamic escape). For Ne, the current ratio of $^{22}Ne/^{20}Ne$ represents a balance between loss to space and continued resupply from the interior; the similarity of the ratio to the terrestrial value is coincidental. For N, the loss by both sputtering and photochemical 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.

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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 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.

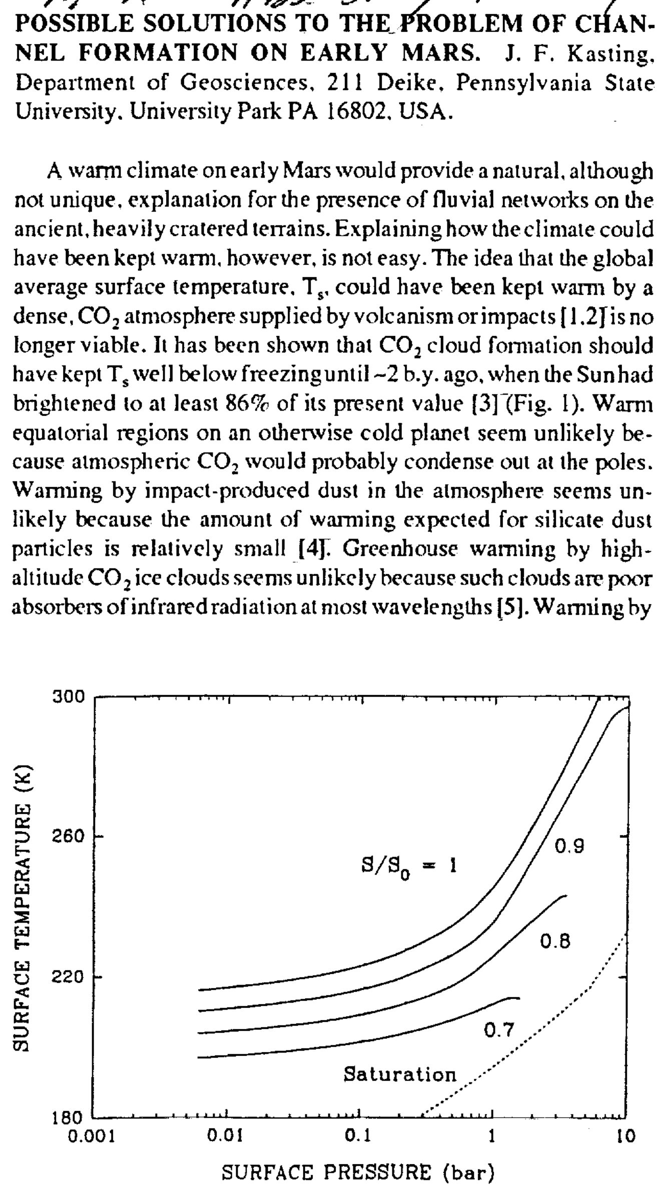

Fig. 1. Mean global surface temperature on Mars as a function of atmospheric CO_2 partial pressure. S/S_o represents the magnitude of the solar luminosity compared to its present value. Solutions with mean surface temperature >273 K are found only for $S/S_o > 0.86$. The dashed curve is the saturation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

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/m³ 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/m³, respectively, when $\upsilon_\infty = 0$, 10, and 20 km/s. Thus, the accretion energy flux ≥200 W/m² 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 ≥10^{18} kg and impact velocity ≥7.25 km/s are physically plausible [5]. Thus, the formation of such magma

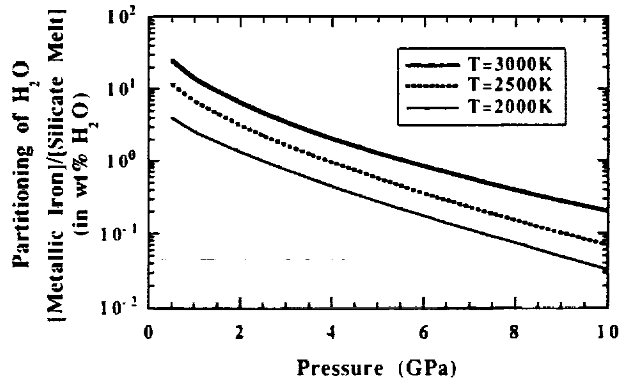

Fig. 1. H_2O partitioning between the silicate melt and liquid metallic Fe. Hydrogen concentration in metallic Fe is shown as a value converted to equivalent H_2O concentration. H_2O fugacity in silicate melt is calculated from the result by [9]. H_2 fugacity in liquid Fe is calculated from the model 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.

ponds by large impacts may be the most plausible mechanism to explain the early formation of the core in Mars.

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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EARLY MARS: A REGIONAL ASSESSMENT OF DENUDATION CHRONOLOGY.

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Within the oldest highland units on Mars, the record of crater degradation indicates that fluvial resurfacing was responsible for modifying the Noachian through middle-Hesperian crater population [1]. Based on crater frequency in the Noachian cratered terrain, age/elevation relations suggest that the highest exposures of Noachian dissected and plateau units became stabilized first, followed by successively lower units [2]. In addition, studies of drainage networks indicate that the frequency of Noachian channels is greatest at high elevations [3]. Together, these observations provide strong evidence of atmospheric involvement in volatile recycling. The long time period of crater modification also suggests that dendritic highland drainage was not simply the result of sapping by release of juvenile water, because the varied geologic units as well as the elevation dependence of stability ages makes it unlikely that subsurface recycling could provide a continuous supply of water for channel formation by sapping. While such geomorphic constraints on volatile history have been established by crater counts and stratigraphic relations using the 1:2M photomosaic series, photogeologic age relationships at the detailed level are needed to establish a specific chronology of erosion and sedimentation. Age relations for discrete erosional slopes and depositional basins will help refine ages of fluvial degradation, assess effectiveness of aeolian processes, and provide a regional chronology of fluvial events. In particular, are stratigraphic relations between dissected plateau units and neighboring plains (usually lumped on small-scale mapping) consistent with a local source/sink scenario for fluvial deposits? Can age relations be determined for discrete depositional basins [e.g., 4] and their neighboring eroded highlands? Did individual degradation events last long enough to be resolved by the cratering record?

One of the long-standing problems in martian geomorphology has been the unique identification of fluvial deposits either within the northern plains [5,6] or elsewhere in the debouchment regions of channels. However, in the low-latitude highlands, sedimentary deposits occur within enclosed basins [4], at areas of channel constriction, and within impact craters. These materials typically consist of subdued, polygonal mesas 2–10 km across, are morphologically similar to the fretted terrain of the Nilosyrtis Mensae region [7], and are laterally confined. Where these deposits are present within degraded craters, the host crater is typically breached by either a through-going or terminating channel. Unfortunately, these units are not extensive enough to allow crater age determinations, so their ages must be inferred by stratigraphy and the age of the superposed surface. Later periods of volcanism and airfall deposition [8] have probably buried many of these deposits, but their distribution suggests that the original sedimentary cover of the martian highlands was once more extensive than is now represented by the few scattered outliers.

In contrast to depositional surfaces, erosional surfaces in the highlands are much more easy to date. There the record of degraded craters indicates the combined effects of erosion from the Noachian through mid Hesperian. The fresh crater population can be used to tell when such surfaces were no longer subject to earlier intense erosion. In the absence of discrete, datable deposits, such erosion surfaces are being used to determine the timing of Mars denudation.

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EARLY MARS WAS WET BUT NOT WARM: EROSION, FLUVIAL FEATURES, LIQUID WATER HABITATS, AND LIFE BELOW FREEZING.

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There is considerable evidence that Mars had liquid water early in its history and possibly at recurrent intervals. It has generally been assumed that this implied that the climate was warmer as a result of a thicker CO_2 atmosphere than at the present. However, recent models suggest that Mars may have had a thick atmosphere but may not have experienced mean annual temperatures above freezing. In this paper we report on models of liquid water formation and maintenance under temperatures well below freezing.

Our studies are based on work in the north and south polar regions of Earth. Our results suggest that early Mars did have a thick atmosphere but precipitation and hence erosion was rare. Transient liquid water, formed under temperature extremes and maintained under thick ice covers, could account for the observed fluvial features. The main difference between the present climate and the early climate was that the total surface pressure was well above the triple point of water.

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WET INSIDE AND OUT? CONSTRAINTS ON WATER IN THE MARTIAN MANTLE AND ON OUTGASSED WATER, BASED ON MELT INCLUSIONS IN SNC METEORITES.

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Constraints on the volatile inventory and outgassing history of Mars are critical to understanding the origin of ancient valley systems and paleoclimates. Planetary accretion models for Mars allow either a volatile-rich [1] or volatile-poor [2] mantle, depending on whether the accreted materials were fully oxidized or whether accretion was homogeneous so that water was lost through reaction with metallic iron. The amount of water that has been outgassed from the interior is likewise a contentious subject, and estimates of globally distributed water based on various geochemical and geological measurements vary from a few meters to more than a thousand meters [3]. New data on SNC meteorites, which are thought to be martian igneous rocks [4], provide constraints on both mantle and outgassed water [5].

The bulk water contents of SNC meteorites, measured after precombustion to remove terrestrial contaminants, are small, in the range of 130–350 ppm. However, because of low internal pressures on Mars, ascending magmas are subject to vesiculation, and they would further desiccate on eruption and contact with the dry atmosphere. Also, the O isotopic composition of water released at high temperature from SNC meteorites suggests that some fraction of the water they contain is not magmatic, but is due to alteration by water already in the crust [6]. Thus bulk water contents are of little use in assessing how much water has actually been delivered to the surface. What is needed is an estimate of the water contents of SNC magmas prior to near-surface degassing and interaction with crustal water.

Many SNC meteorites contain crystals formed at depth and the cores of these crystals contain now-solidified pockets of trapped melt. Some melt inclusions contain daughter crystals of hydrous amphibole (kaersutite), a phase that does not appear in these meteorites outside the inclusions. Phase equilibria for kaersutite [7] indicate that it is only stable at pressures above 1.5 kbar, corresponding to depths on Mars of 11 km. This is below the self-compression depth for martian crust, and magma trapped at such depth is unlikely to have experienced vesiculation or interaction with crustal water. Kaersutite crystallizes only when the water contents of inclusion melts reach 4 wt% [7]. From the extent of inclusion solidification before the onset of kaersutite crystallization, we can estimate the amount of water in the magma at the time of trapping. The solidification histories for melt inclusions from SNC meteorites have been modeled using linear regression methods to solve a system of mass-balance equations [5,7]. The results for SNC melt inclusions typically indicate 50–75% crystallization before kaersutite forms, corresponding to initial water contents in these magmas of approximately 1.4 wt%.

If we assume that SNC magmas are representative of martian volcanism, we can combine this water content with visual estimates of the total volume of martian igneous materials [8] to obtain an outgassed water depth of approximately 200 m. This water estimate also rests on the assumption that intrusions in the subsurface effectively degassed, and may serve as a refined lower limit for martian water outgassed since 3.9 b.y. ago. Geological evidence for greater amounts of surface water [3,9] would then imply significant outgassing before formation of a stable crust or heterogenous accretion of a veneer of cometary matter.

These data also have implications for the water content of the martian interior. The 36 ppm water for the mantle estimated from the geochemical model of [2] seems too low because it would imply an implausibly small degree of melting (0.2%, if water is perfectly incompatible and no subsequent fractionation of the magma occurred) to produce SNC magmas. For a more reasonable 10% melting, a magma with 1.4% water requires a source region containing 1400 ppm water. This value is too high because SNC meteorites clearly formed from fractionated magmas, but it does suggest that the martian mantle is signficantly wetter than has been inferred from some past geochemical and geologic models [9], perhaps having a similar water content to the terrestrial mantle. This suggestion is not really surprising considering the enrichment of other volatiles in Mars [2], but it may require that water reacted incompletely with Fe during core formation. The conclusion that the martian mantle may have been quite wet (even as late as 1.3 b.y. ago, the age of SNC meteorites) is important in assessing its capacity to outgas significant amounts of water prior to 3.9 b.y. ago.

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THE YOUNG SUN AND PHOTOCHEMISTRY OF THE PRIMITIVE MARTIAN ATMOSPHERE. H. Nair, M. F. Gerstell, and Y. L. Yung, Division of Geological and Planetary Sciences 170-25, California Institute of Technology, Pasadena CA 91125, USA.

Many investigators of the early martian climate have suggested that a dense CO_2 atmosphere was present in order to warm the surface above the melting point of water [e.g., 1]. However, Kasting [2] recently pointed out that previous thermal models of the primitive martian atmosphere had not considered the condensation of CO_2. When this effect was incorporated, Kasting found that a purely CO_2 greenhouse is an inadequate mechanism to warm the surface.

Observations of young stars, both premain sequence and early main sequence, indicate that their ultraviolet luminosities are much higher than the present ultraviolet output of the Sun. If such behavior is a normal phase of stellar evolution, we may expect that the Sun also had a substantially enhanced ultraviolet luminosity in its youth [3]. This has significant implications for the martian atmosphere as CO_2 is rapidly dissociated by ultraviolet photons shortward of 2000 Å.

Our photochemical model shows that under the influence of the early solar ultraviolet spectrum, an initial reservoir of CO_2 is decomposed to the extent that CO and O_2 become the major components of the atmosphere. Large ozone densities arise due to the increased O_2 abundance. Similar investigations for the early terrestrial atmosphere have also shown that CO, O_2, and O_3 concentrations are markedly enhanced when the model atmosphere is subjected to more intense ultraviolet fluxes [4,5].

We will investigate the climatology of an atmosphere where CO_2 is a minor constituent but still the key radiative species. The thermal structure of the dust-free atmosphere is estimated by employing a simple radiative-convective model similar to that used by Gierasch and Goody [6]. Radiative heating rates are computed using the Caltech/JPL one-dimensional photochemical model. Thermal cooling rates for a martian atmosphere containing O_2, O_3, H_2O, N_2O, CO, and CO_2 are calculated using FASCODE [7] and k-distribution methods [8]. The effects due to pressure broadening of the infrared absorption lines of CO_2 by CO and O_2, as well as the radiative effects of increased ozone densities in the atmosphere, will be examined.

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S21-91 ABS. N94-21680 !

p. 1

EVOLUTION OF THE MARTIAN ATMOSPHERE. R. O. Pepin, School of Physics and Astronomy, University of Minnesota, Minneapolis MN 55455, USA.

Evolution of Mars' noble gases through two stages of hydrodynamic escape early in planetary history has been proposed previously by the author [1]. In the first evolutionary stage of this earlier model, beginning at a solar age of ~50 m.y., fractionating escape of a H_2-rich primordial atmosphere containing CO_2, N_2, and the noble gases in roughly the proportions found in primitive carbonaceous (CI) chondrites is driven by intense extreme-ultraviolet (EUV) radiation from the young evolving Sun. Hydrogen exhaustion then leads to a long (~80 m.y.) period of quiescence, followed by abrupt degassing of remnant H_2, CO_2, and N_2 from the mantle and of solar-composition noble gases lighter than Xe from the planet's volatile-rich accretional core. Degassed H refuels hydrodynamic loss in a waning but still potent solar EUV flux. Atmospheric Xe, Kr, and Ar remaining at the end of this second escape stage, ~4.2 G.y. ago, have evolved to their present-day abundances and compositions. Residual Ne continues to be modified by accretion of solar wind gases throughout the later history of the planet.

This model does not address a number of processes that now appear germane to martian atmospheric history. One, gas loss and fractionation by sputtering, has recently been shown to be relevant [2,3]. Another, atmospheric erosion, appears increasingly important [4–6]. In the absence then of a plausible mechanism, the model did not consider the possibility of isotopic evolution of noble gases heavier than Ne after the termination of hydrodynamic escape. Subsequent nonthermal loss of N [7] was assumed, in an unspecified way, to account for the elevation of $\delta^{15}N$ from the model value of ~250‰ at the end of the second escape stage to ~620‰ today. Only qualitative attention was paid to the eroding effects of impact on abundances of all atmophilic species prior to the end of heavy bombardment ~3.8 G.y. ago. No attempt was made to include precipitation and recycling of carbonates [8] in tracking the pressure and isotopic history of CO_2.

All these evolutionary processes, and others, can in fact be modeled in a straightforward way along with hydrodynamic escape. However, their inclusion requires a different mathematical architecture than the closed-form integration of analytic equations across entire escape episodes utilized in [1] to determine the effects of hydrodynamic loss acting alone. An approach in which each of several mechanisms operates independently over short time intervals serves very well [3,9], although at the cost of some computational complexity. In this approach martian atmospheric history is divided into small timesteps, Δt, in the present model of average duration ~0.5 m.y. and ~4 m.y. respectively for times earlier and later than 3.8 G.y. ago. Evolutionary tracking begins ~4.5 G.y. ago at a solar age of ~100 m.y., when a H-rich primordial atmosphere containing CO_2, N, and noble gases of mixed CI-solar composition, degassed by impact from accreting meteoritic and cometary planetesimals during planetary growth, is presumed to surround the planet [1]. The first and subsequent timesteps include evolution from initial atmospheric abundances and isotopic compositions by whichever of the following loss and addition mechanisms are judged to be operative during that interval: EUV-driven hydrodynamic escape, atmospheric erosion by impact, planetary outgassing, sputtering from the exobase by exospheric "pick-up" ions, photochemical escape (for N), and carbonate formation and recycling (for CO_2).

Each of these processes is assumed to act independently on the volatile inventories present at the beginning of each Δt timestep. Initial abundances and isotopic compositions for the following timestep are adjusted to reflect losses, gains, and isotopic shifts generated in the atmospheric and carbonate reservoirs during the preceding interval.

This more general procedure has been used to track the noble gases, CO_2, and N from primordial inventories to their present compositional states in a revised model of atmospheric evolution on Mars [9]. Atmospheric history is divided into early and late evolutionary periods, the first characterized by high CO_2 pressures and a possible greenhouse [8] and the second by a low-pressure cap-regolith buffered system [10] initiated by polar CO_2 condensation [11], assumed for illustration to have occurred ~3.8 G.y. ago. During early evolution the Xe isotopes are fractionated to their present composition by hydrodynamic escape, and CO_2 pressure and isotopic history is dictated by the interplay of losses to erosion, sputtering, and carbonate precipitation, additions by outgassing and carbonate recycling, and perhaps also by feedback stabilization under greenhouse conditions. Atmospheric collapse leads to abrupt increases in the mixing ratios of preexisting Ar, Ne, and N_2 at the exobase and their rapid removal by sputtering [3]. Current abundances and isotopic compositions of these light species are therefore entirely determined by the action of sputtering and photochemical escape on gases supplied by planetary outgassing during the late evolutionary epoch. The present atmospheric Kr inventory also derives almost completely from solarlike Kr degassed during this period. Consequently, among current observables, only the Xe isotopes and $\delta^{13}C$ survive as isotopic tracers of atmospheric history prior to its transition to low pressure. With the possible exception of $\delta^{13}C$, this baseline model generates very satisfactory matches to current atmospheric abundances and isotopic compositions.

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EARLY MARS: THE INEXTRICABLE LINK BETWEEN INTERNAL AND EXTERNAL INFLUENCES ON VALLEY NETWORK FORMATION. S. E. Postawko[1] and F. P. Fanale[2], [1]School of Meteorology, University of Oklahoma, Norman OK 73019, USA, [2]Planetary Geosciences, University of Hawaii, Honolulu HI 96822, USA.

The conditions under which the valley networks on the ancient cratered terrain on Mars formed are still highly debated within the scientific community. While liquid water was almost certainly involved (although this has recently been questioned [1]), the exact mechanism of formation is uncertain. The networks most resemble terrestrial sapping channels [2], although some systems exhibit a runoff-dominated morphology [3]. The major question in the fomation of these networks is what, if anything, do they imply about early martian climate?

There are typically two major theories advanced to explain the presence of these networks. The first is that higher internal regolith temperatures, associated with a much higher heat flow 3.8 b.y. ago, would cause groundwater to be closer to the surface than at present [4]. Just how close to the surface groundwater would have to exist in order to form these valley networks has recently been questioned [3]. The second major theory is that early Mars had a much thicker atmosphere than at present, and an enhanced atmospheric greenhouse may have increased surface temperatures to near the freezing point of water [5–7]. While recent calculations indicate that CO_2 alone could not have produced the needed warming [8], the presence of other greenhouse gases [8–10] may have contributed to surface warming.

It does not, in fact, make sense on physical grounds to consider these two mechanisms separately. The effectiveness of both atmospheric greenhouse warming and higher internal regolith temperatures on early Mars is dependent on high early heat flow. In the case of the atmospheric greenhouse effect, this is because the abundance of a greenhouse gas (or any gas) in the atmosphere will depend on the supply rate from the interior (which may include supply of both juvenile and recycled gases), which can be related to the heat flow. The depth to the liquid water level, which depends on internal regolith temperatures, can also be related to heat flow. We have derived a quantitative relationship between the effectiveness of an atmospheric greenhouse and that of internal regolith temperature in producing the morphological differences between early and later martian terrains. While our arguments here are based on CO_2 as the dominant atmospheric gas, similar arguments can undoubtedly be made for the supply of other gases to the atmosphere.

For any set of martian orbital parameters and level of solar activity, the atmospheric CO_2 pressure controls the surface temperature. For a chosen total CO_2 inventory (atmosphere plus regolith), and specified atmospheric mean residence time for CO_2, the CO_2 atmospheric pressure is controlled by the mean residence time of CO_2 in the regolith as carbonate. It has been shown [7] that the atmospheric P_{CO_2} can be expressed as a function of heat flow. In addition, for any given regolith conductivity, the heat flow equation also allows temperature at any depth (and thus the depth to the 273 K isotherm) to be expressed as a function of surface temperature and heat flow. Therefore, for any assumed atmospheric mean residence time, regolith conductivity, and total available CO_2 inventory, the depth to the 273 K isotherm can be expressed as a function of surface temperature. The relationship between surface temperature and depth to the 273 K isotherm has been derived using relationships already in the literature [7,8]. Figure 1 illustrates that, for a given set of assumptions, the relative roles of internal regolith temperature and atmospheric greenhouse effect are inextricably interlocked. In all cases, the atmospheric mean residence time of CO_2 is taken to be 10^7 yr, and regolith conductivity is 0.5×10^3 mW m^{-1} K^{-1}. The numbers in parentheses correspond to heat flow, in units of mW m^{-2}. In Fig. 1a the total available CO_2 inventory is 1 bar, while in Fig. 1b the total CO_2 inventory is 5 bar. It is clear from these figures that if the total available CO_2 was on the order of 1 bar, then the atmospheric greenhouse effect plays a very minor role in raising the depth at which liquid water may be found, particularly in areas outside the equatorial region. If the total available CO_2 was at the upper end of the expected value range (4 bar or more), then the atmospheric greenhouse effect dominates almost completely in the equatorial region. However, on a global basis the suppression of

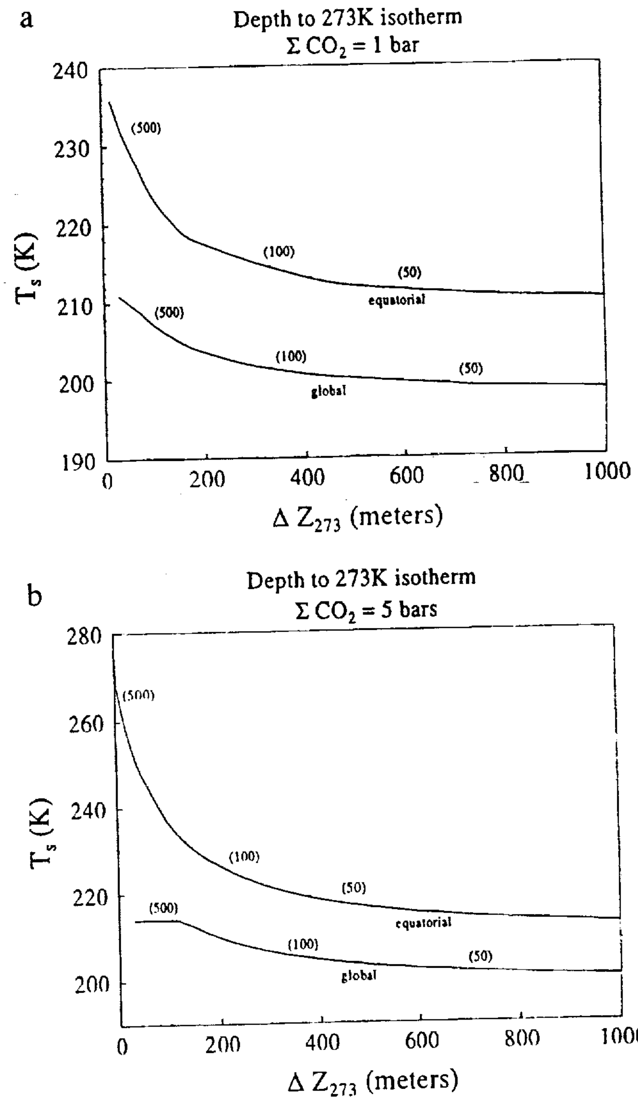

Fig. 1.

surface warming by condensation of atmospheric CO_2 [8] means that even when relatively large amounts of CO_2 are available the atmospheric greenhouse effect is minimal. Nevertheless, even in these cases, a change in internal gradient accompanying a higher early heat flow can still easily decrease the depth to liquid water from well over a kilometer to less than 350 m.

Many uncertainties still exist in assessing the absolute importance of an atmospheric greenhouse effect vs. higher internal regolith temperatures on early Mars. These include the question of regolith conductivity and the magnitude and timing of changes in solar luminosity, as well as the depth at which liquid water must exist in order to be able to form the valley networks. In addition, although we have examined the effectiveness of atmospheric greenhouse warming due to CO_2, as was previously noted it is possible that other greenhouse gases were present in the early martian atmosphere. Any calculations of the greenhouse contribution of these gases, some of which may have had relatively short atmospheric lifetimes, will need to take into account the rate at which they may have been supplied to the atmosphere. Schemes analogous to that presented for CO_2 will have to be explored in order to assess the absolute contribution of any potential greenhouse gas on early Mars.

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THE MARTIAN VALLEY NETWORKS: ORIGIN BY NIVEO-FLUVIAL PROCESSES. J. W. Rice Jr., Department of Geography, Arizona State University, Tempe AZ 85287, USA.

The valley networks may hold the key to unlocking the paleoclimatic history of Mars. These enigmatic landforms may be regarded as the martian equivalent of the Rosetta Stone. Therefore, a more thorough understanding of their origin and evolution is required. However, there is still no consensus among investigators regarding the formation (runoff vs. sapping) of these features.

Recent climatic modeling [1] precludes warm (0°C) globally averaged surface temperatures prior to 2 b.y. when solar luminosity was 25–30% less than present levels. This paper advocates snowmelt as the dominant process responsible for the formation of the dendritic valley networks. Evidence for martian snowfall and subsequent melt has been discussed in previous studies. It has been suggested [2] that Mars has undergone periods of very high obliquities, up to 45°, thus allowing snow accumulations, several tens of meters thick, at low latitudes as a result of sublimation from the poles. Clow investigated the conditions under which snow could have melted by solar radiation by using an optical-thermal model developed for dusty snowpacks [3]. It was found that the low thermal conductivity of snow and its partial transparency to solar radiation can result in subsurface melting despite surface temperatures well below freezing. Melting and subsequent runoff can occur at atmospheric pressures as low as 30–100 mbar [3]. Carr showed that if streams 2 m deep or larger can be initiated and sustained, then flows up to a few hundred kilometers long can be established, even under present-day climatic conditions [4]. Therefore, based on the above-mentioned work, it seems logical to the author that snowfall and subsequent snowmelt has many advantages to other explanations for the formation of the valley networks.

It has been argued that the valley networks were formed primarily by groundwater seepage. This is based on the measurement of junction angles between intersecting tributaries and on morphologic characteristics that appear to suggest headward extension through basal sapping [5]. The evidence for sapping is in some cases convincing (i.e., Nirgal Vallis), but it does not explain many of the dendritic valley systems, e.g., those located in the Margaritifer Sinus region.

Some problems with the sapping model will be discussed below. First, the measurement of junction angles between individual intersecting tributaries of the valley networks does not provide evidence to refute the view that the networks were formed by rainfall/snowmelt-fed erosion. Stream junction angles are controlled by slope, structure, lithology, and basin development stage, not precipitation [6]. Sapping requires that zones of low hydraulic head somehow be established to support the gradients needed to allow groundwater flow, and that zones of high hydraulic head be recharged, presumably by precipitation. Additionally, some of the valley networks whose channels originate on crater-rim crests indicate that the local water table must have intersected the surface high on the crater wall if sapping was involved [3]. This would mean that the crater was once filled with water, but there is no evidence, such as inflowing channels, to support this condition. It should also be noted that all the valley networks have been modified by mass wasting processes such as gelifluction and thermal erosion.

In order to more fully understand niveo-fluvial systems on Mars one should study terrestrial periglacial regions such as the Northwest Territories in the Canadian High Arctic. It is proposed that the following geomorphic processes and resulting landforms of snowmelt-fed rivers be used to explain the dendritic valley networks on Mars.

The Mecham River near Resolute, Northwest Territories, provides an excellent example of stream action and valley development in the periglacial realm. The area is underlain by continuous permafrost and mean monthly air temperatures are below zero for 9–10 months a year. The Mecham River has 80–90% of its annual flow concentrated in a 10-day period. This is typical for periglacial rivers in the High Arctic. During this brief period of concentrated flow extensive movement of bedload occurs, sometimes with peak velocities up to 4 m/s, causing the whole bed to be in motion [7]. This pattern of intense activity has far greater erosive and transporting potential than a regime in which river flow is evenly distributed throughout the year. The dominance of bedload movement in Arctic streams helps explain the distinctive flat-bottomed form of many periglacial stream valleys [8]. Thermal erosion and the subsequent collapse of river banks provides material for bedload transport and deposition downstream. This process also aids in the development of the broad flat-floored valleys. The permafrost also favors the flat-floored valley profiles because it provides a near-surface limit to downward percolation of water, thereby promoting runoff [9]. Another interesting feature of these periglacial rivers is that they lack a pronounced channel on their floors. This holds true for valleys eroded into either bedrock or unconsolidated debris.

Other work [10] indicates that fluvial processes have often been underestimated in periglacial regions. Budel illustrates this point in Spitsbergen, where he pointed out that ground ice breaks apart the rocks and prepares them for fluvial action. Periglacial rivers do not need to carry out new erosive action but need only melt the eisrinde and transport the shattered debris. The eisrinde is composed of the upper frozen and highly shattered layer of the permafrost. Rivers operating under this regime can deepen their beds rapidly; downcutting rates on the order of 1–3 m/1000 yr over the last 10,000 yr have been estimated for Spitsbergen [10].

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MARS AND THE EARLY SUN. D. P. Whitmire[1], L. R. Doyle[2], R. T. Reynolds[3], and P. G. Whitman[1], [1]University of Southwestern Louisiana, Lafayette LA 70504-4210, USA, [2]SETI Institute, NASA Ames Research Center, Moffett Field CA 94035, USA, [3]Theoretical Studies, NASA Ames Research Center, Moffett Field CA 94035, USA.

Global mean temperatures near 273 K on early Mars are difficult to explain in the context of standard solar evolution models. Even assuming maximum CO_2 greenhouse warming, the required flux is ~15% too low [1]. Here we consider two astrophysical models that could increase the flux by this amount. The first model is a nonstandard solar model in which the early Sun had a mass somewhat greater than today's mass (1.02–1.06 $M_\odot$). The second model is based on a standard evolutionary solar model, but the ecliptic flux is increased due to focusing by an (expected) heavily spotted early Sun.

The relation between stellar mass M and luminosity L for stars near 1 $M_\odot$ is $L \sim M^{4.75}$ [2]. If the Sun's original mass were larger than at present, the early planetary flux would be further increased due to migration of orbits. Isotropic mass loss does not produce a torque on a planet and so angular momentum is conserved. Consequently, semimajor axes increase inversely with mass loss and the flux is proportional to $M^{6.75}$. To increase the flux at Mars by 15% requires that the Sun's mass be $\geq 1.02\ M_\odot$. On the other hand, the flux cannot be so large (1.1× that of the flux at 1 AU today) that Earth would have lost its water [3]. This imposes an upper mass limit of 1.06 $M_\odot$.

Nonclimatic evidence for mass loss of this magnitude might be found in the ion implantation record of meteorites and Moon rocks. Such evidence does exist, but is inconclusive due to uncertainties in exposure times and dating [4,5]. The dynamical record of adiabatic mass loss is also inconclusive. The adiabatic invariance of the action variables implies that the eccentricities and inclinations of planetary orbits remain constant as the semimajor axes increase. The dynamical drag of the wind would have no effect on planets, but would cause a net inward migration of bodies of sizes less than about 1 km [6]. Whether the cratering record is consistent with this dynamical consequence is unclear. Mass loss could also be an additional process contributing to bringing organics into the inner solar system.

A mass loss of 0.1 $M_\odot$ has been suggested as an explanation for the depletion of Li in the Sun by 2 orders of magnitude over primordial values [7]. However, this explanation has been reconsidered by [8], who find that mass loss cannot explain the depletion of Li in Hyades G dwarfs. Although it is generally believed that young G stars are spun down by mass loss, most models are insensitive to the total mass loss required [e.g., 9]: An exception is the model by [10] which predicts a mass loss comparable to our lower limit.

The most promising nonclimatic evidence for main sequence mass loss from the early Sun is the direct observation of similar mass loss from young main-sequence G stars. Detection of stellar mass loss from late dwarfs at the predicted rate (less than $\sim 10^{-10}\ M_\odot\ yr^{-1}$) by optical techniques is generally not possible. However, in one unique case where it could be measured, an outflow 1000× that of the present Sun was found in a K2V dwarf [11]. Recently, huge winds have been reported from several M dwarfs [12]. This technique involves detections over a wide range of radio and millimeter wavelengths and the fact that free-free emission from an optically thick wind has a characteristic spectrum in which the flux is proportional to the 2/3 power of the frequency. As a first step in extending this technique to solar-type stars we have recently used the VLA to obtain the radio emission at 2 and 6 cm in four nearby young G-type stars.

In the second model (ecliptic focusing) we assume standard solar evolution. Young G stars are often observed to be heavily spotted (10–50%). In contrast to mature G stars like the Sun, which typically have only a maximum coverage of ~0.1%, the net effect of star spots on young G stars is to reduce the radiated flux at the location of the spot. Since the total stellar luminosity is determined by nuclear reactions in the core, the flux must increase in regions without spots. Such variations in flux are observed on short (days) and long (years) timescales [13]. These observations measure the anisotropy in the distribution of spots. A more significant effect would be the average increase in the equatorial flux if the time-averaged location of the spotted regions was nearer to the stellar poles than to the equator. This is not the case in today's Sun, but is observed to occur in young stars such as the G2V star SV Camelopardalis, in which there is a ~10% coverage, localized in latitude and longitude, toward one of the poles.

We have investigated a simple model in which polar cap blocking focuses the stellar flux in the equatorial plane. The equatorial flux can be enhanced a maximum of a factor of 2 over the uncapped case. For a time-averaged polar coverage of 10% the equatorial flux enhancement factor is 1.17. Refinements in this model and a review of the relevant observational data will be presented.

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