Basal melting of snow on early Mars: A possible origin of some valley networks

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[1] Valley networks appear to be cut by liquid water, yet simulations suggest that early Mars could not have been warmed enough by a CO_2-H_2O greenhouse to permit rainfall. The vulnerability of an early atmosphere to impact erosion, the likely rapid scavenging of CO_2 from the atmosphere by weathering, and the lack of detection of weathering products all support a cold early Mars. We explore the hypothesis that valley networks could have formed as a result of basal melting of thick snow and ice deposits. Depending on the heat flow, an early snowpack a few hundred meters to a few kilometers thick could undergo basal melting, providing water to cut valley networks. *INDEX TERMS:* 5415 Planetology: Solid Surface Planets: Erosion and weathering; 5418 Planetology: Solid Surface Planets: Heat flow; 6225 Planetology: Solar System Objects: Mars. **Citation:** Carr, M. H., and J. W. Head III, Basal melting of snow on early Mars: A possible origin of some valley networks, *Geophys. Res. Lett.*, *30*(24), 2245, doi:10.1029/2003GL018575, 2003.

1. Introduction

[2] Much of the ancient cratered surface of Mars is dissected by branching valley networks that superficially resemble terrestrial drainage systems. The valleys have been taken as evidence of former warm, wet conditions particularly early in the planet's history when most of the valley networks formed (summarized in *Baker*, 2001; *Carr*, 1996; *Craddock and Howard*, 2002]. The perception of an early, warm Mars is, however, being increasingly questioned because of (1) failure to detect weathering products from orbit [*Christensen, et al.*, 2001], (2) the vulnerability of an early atmosphere to losses by impact erosion [*Melosh and Vickery*, 1989], (3) the likely rapid scavenging of CO_2 from the atmosphere by weathering under warm, wet conditions [*Pollack et al.*, 1987], and (4) climate modeling studies which show that it is difficult, if not impossible, to sufficiently warm Mars with a CO_2-H_2O greenhouse so that rainfall could occur [*Haberle*, 1998; *Kasting*, 1991]. The latter problem is particularly acute early in the planet's history, when the Sun's output was likely significantly less than it is today [*Newman and Rood*, 1987]. Geologic studies have shown that many of the valley networks have characteristics suggestive of both groundwater sapping [e.g., *Pieri*, 1980; *Baker*, 1990] and surface runoff [e.g., *Craddock and Howard*, 2002; *Hynek and Phillips*, 2001]. Sustained groundwater sapping requires some form of groundwater recharge. While recharge may be accomplished in part by basal melting of the south polar cap [*Clifford*, 1987], this cannot be the only mechanism for groundwater recharge because large numbers of the valleys occur at elevations well in excess of the base of the south polar cap [*Carr*, 2002]. If the valleys were cut by water, as appears likely, some form of precipitation is required. If the climate modeling studies are correct, precipitation of rain is highly improbable because global temperatures remain well below freezing for a CO_2-H_2O greenhouse no matter how thick the atmosphere [*Haberle*, 1998]. Possible exceptions are during short anomalous periods immediately following large impacts [*Segura et al.*, 2002]. Here we explore the possibility that the valley networks formed under cold conditions by the basal melting of snow, which could have provided water both to directly erode the valleys and to recharge the groundwater system.

[3] Widespread water ice deposits may have been present on early Mars because of (1) the likely cold temperatures, (2) the large near surface inventory of water, as indicated by the amount of erosion needed to cut the valley networks, and subsequently the outflow channels, (3) higher heat flow levels that would cause most of the near surface inventory of water to be on the surface, rather than incorporated into the cryosphere [*Clifford and Parker*, 2001]. In addition, *Jakosky and Carr* [1985] suggested that snow would be precipitated at low latitudes during periods of high obliquity, a suggestion now supported by GCM simulations [e.g., *Mischna et al.*, 2003].

[4] If a snowpack were present at low latitudes, meltwater could be derived by melting from above by sunlight or by melting from below by internal heat [*Zent*, 1999]. *Clow* [1987] explored the first possibility and found that under martian conditions, if the snowpack contained the appropriate amount of dust, small amounts of meltwater could be generated. However the amounts are small, and while melting of a snowpack by sunlight may explain recent gullies [*Costard et al.*, 2002], it is unlikely to generate enough water to collect into streams large enough that they could flow for hundreds to thousands of kilometers, as is required to explain the valley networks. One problem with the solar melting scenario is that the meltwater generated at the top of the snowpack tends to re-freeze as it trickles downward. In contrast, if internal heat were responsible for the melting, meltwater could slowly accumulate at the base of the snowpack and either infiltrate into the warm ground,

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collect into sub-snowpack streams, or be released episodically. If this mechanism was the source of water that cut the valleys and recharged the groundwater system, then the rapid decline in the rate of valley formation at the end of the Noachian is readily explained. It simply results from the declining heat flow.

2. Snowpack Densities and Thermal Conductivities

[5] A snowpack on Mars can be compared to the dry snow zones of Greenland and Antarctica, where mean annual surface temperatures are 248 K or less [*Paterson*, 1994], and the density of the snow increases with depth as a result of self-compaction and pressure sintering. An empirical density-depth relation is

$$\rho = \rho_i - (\rho_i - \rho_s)\exp(-Cz) \quad (1)$$

where ρ is the density at depth z, ρ_i is the density of ice (917 kg m^{-3}) and ρ_s is the surface density. The value of C is typically in the range of 0.02 to 0.03 (SI). Because of the lower gravity on Mars the rate of compaction with depth will be lower. The depth-density relation for martian snow can be obtained by scaling C to Mars' gravity. Alternatively, we can assume that the relationship between density and overburden pressure for Antarctic dry zone ice [*Paterson*, 1994, p. 17] applies to Mars and determine the density-depth profile under martian gravity.

[6] To determine the temperature profile in the snow and the depth to melting we need to know the thermal conductivity of snow and how it varies with depth. Most thermal conductivity measurements of snow have been performed within a few degrees of freezing under a terrestrial atmosphere (summarized in Figure 1, adapted from *Mellor* [1977]). Because of the desire to take into account martian conditions, for which no empirical data were available, *Clow* [1987] estimated thermal conductivities theoretically as follows. The effective thermal conductivity of the snow depends on K_i, the conductivity of ice and K_p, the conductivity of the pores.

$$K_i = 9.28\exp(-0.0057T)\ \mathrm{W\ m^{-1}K^{-1}} \quad (2)$$

K_p is the sum of the conductivity of the gas in the pores (Kg), given by

$$Kg = \rho^* C_p D^* (T/T^*)^{n-1}\,\mathrm{W\ m^{-1}K^{-1}} \quad (3)$$

and the conductivity due to water vapor diffusion (K_v) given by

$$K_v = 4D_v^*(\mu L_s/RT^*)^2(P^*/P)e_s/T\ \mathrm{W\ m^{-1}K^{-1}} \quad (4)$$

An additional term, radiation transfer across the pores, was ignored here because of the low martian temperatures. In the above equations ρ^* is the density of CO_2 under standard temperature (T*) and pressure (P*). Cp and D_h^* are the specific heat and diffusivity constant respectively of CO_2 and n has a value of 2. D_v^* is the molecular diffusivity of water through CO_2, μ is the molecular weight of water, R is the gas constant, L_s is the latent heat of sublimation of water, e_s is the saturation vapor pressure of water over ice.

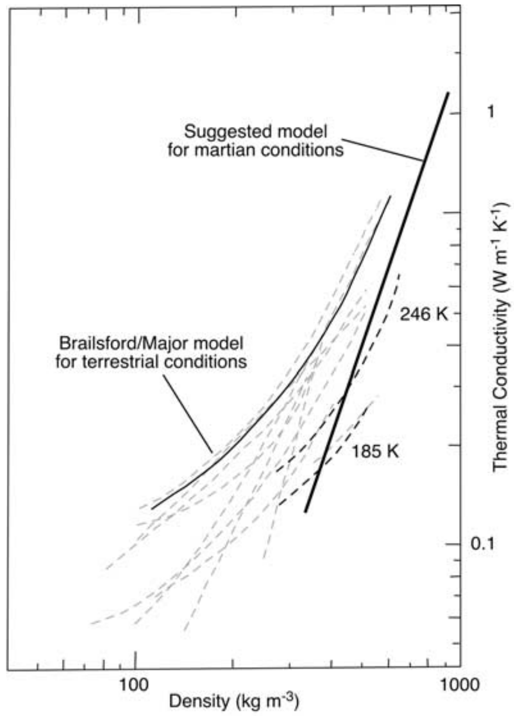

Figure 1. Thermal conductivity vs. snow density. The faint lines are terrestrial experimental data as summarized in *Mellor* [1977]. The effective thermal conductivity estimated by the *Brailsford and Major* [1964] model represents an upper bound. Martian conductivities should be lower than terrestrial experimental values because of lower martian temperatures. The effect of temperature is shown by the 246 K and 185 K curves from *Pitman and Zuckerman* [1967]. The preferred model for Mars (equation 6) extends from the value for pure ice, through the 246 K data, along the lower bound of the terrestrial data.

[7] The effective thermal conductivity of the snow depends on combining the ice conductivity (K_i) and the pore conductivity (K_p). For temperatures below 250 K the pore conductivities are in the 0.01 to 0.1 W m^{-1} K^{-1} range, as compared with the thermal conductivity of ice, which ranges from 2 to 2.5 W m^{-1} K^{-1}. The effective conductivity of the snow thus depends sensitively on how the two conductivities contribute to the total conductivity. *Clow* [1987] found that the *Brailsford and Major* [1964] model for combining K_i and K_p gives values for the effective thermal conductivity of snow close to the upper bound of the measured values given in Figure 1. In this model, the effective thermal conductivity of the snow (K_e) is given by

$$K_e = (K_i/4)\Big[(3\phi-1)(K_p/K_i)+2-3\phi. + \Big\{\big[(3\phi-1)(K_p/K_i)+2-3\phi\big]^2+(8K_p/K_i)\Big\}^{0.5}\Big] \quad (5)$$

where ϕ is volume fraction of pores. Equations (2) through (5) were used to derive an upper bound for martian conditions.

[8] The conductivity of the pores increases with temperature, mostly because of the exponential dependence of the

vapor pressure of water (e_s) on temperature. Since most of the experimental measures of conductivity in Figure 1 were made at temperatures just below 273 K, the values in the figure likely overestimate the pore conductivities for martian temperatures, which typically range from 200–250 K in the critical, upper parts of the profile where the porosity is significant. Measurements by *Pitman and Zuckerman* [1967] at temperatures of 246 K and 185 K fall at or below the rest of the data in Figure 1, as expected from equations (3) and (4). Because of the cold temperatures, the thermal conductivities in the upper part of the Mars profiles are probably close to the 246 K data and at the low side of most of the terrestrial measurements. The expression

$$\log(K_e) = 0.4 + 2.9 \ \log\rho \tag{6}$$

for ρ in the 300 to 917 kg m^{-3} range was adopted here as the best estimate of the martian conductivities since it gives values at the lower bound of the terrestrial data and passes through the 246 K data. An increase in the CO_2 pressure in the pores causes the conductivity of the snow to decrease because of the suppression of water vapor diffusion. In this respect, snow differs from porous rock materials, the conductivity of which decreases with decreasing pore pressure [e.g., *Presley and Christensen*, 1997].

3. Temperature Profiles

[9] Heat flow is the most important factor controlling the temperature profile in the snow and the depth to melting but is the least well-known and most difficult to model. Early models suggested that in the Late Noachian, 3.8–4.0 Gyr ago, mean global heat flow was ~0.15 W m^{-2} [summarized in *Schubert et al.*, 1992]. More recently, various MGS data suggest lower values for heat flow during the Late Noachian. For example, combined MGS gravity and topography data suggest that the preservation of Hellas basin topography places constraints on ductile flow in the crust and thus heat flow [*Zuber et al.*, 2000]. The preservation of distinctive magnetic anomalies in the southern upland crust [*Acuna et al.*, 1999; *Connerney et al.*, 1999] interpreted to be of ancient origin places constraints on Curie point temperature distribution and implies lower crustal heat flux. Localized gravity/topography admittances and correlations in the spectral domain [*McGovern et al.*, 2002] permit estimates of elastic lithospere thickness and derived heat fluxes. These data all show a relatively lower local and global heat flow in the Late Noachian than had been thought in the past from thermal history models. Spacing and nature of wrinkle ridges, many discovered in the northern lowlands by MGS, permit estimates of surface heat flux from a different standpoint [*Montesi and Zuber*, 2003]; these values are broadly consistent with, but slightly higher, than those of *McGovern et al.* [2002]. *Spohn et al.* [2001] updated the thermal history models described in *Schubert et al.* [1992] based on early MGS results and lowered the Late Noachian heat flow estimates into the 0.05 to 0.08 W m^{-2} range, while acknowledging that there is considerable uncertainty in the models. Furthermore, global heat loss trends may be non-linear [*Choblet and Sotin*, 2001] and heterogeneous at any given time [e.g., *McGovern et al.*, 2002]. The results over this range for both a surface temperature of 210 K, close to today's mean surface temperature, and 230 K are shown in Figure 2. The results suggest that for the preferred conductivities, and with heat flow less than 0.1 W m^{-2}, melting temperatures are not reached until depths of several hundred meters or more. For heat flows in the 0.1 to 0.15 W m^{-2} range, melting occurs at depths of a few hundred meters. If Mars had a significantly higher surface inventory of CO_2 at the end of heavy bombardment than at present, a bar or more, higher surface temperatures are possible [*Haberle et al.*, 1994]. Recent spectroscopic identification of carbonate minerals at the 2–5 weight % level in martian dust [*Bandfield et al.*, 2003] have led to the suggestion that these carbonate minerals may be a large sink of a thicker past martian atmosphere. *Bandfield et al.* [2003] estimate that a global layer 1–3 km thick containing 2% carbonate can account for ~1–3 bars of CO_2. Limited losses of CO_2 to space after heavy bombardment [*Jakosky and Jones*, 1997] suggests that such conditions on early Mars should be seriously considered. If surface temperatures were as high as 230 K, melting would occur at snowpack depths between 50 and 200 m for heat flows in the 0.1 to 0.15 W m^{-2} range.

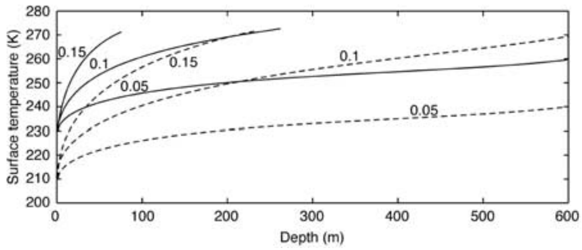

Figure 2. Depth-temperature profiles for a surface temperature of 210 K (dashed lines) and 230 K (solid lines), assuming the preferred thermal conductivity shown in Figure 1. Three curves are shown for each surface temperature corresponding to heat flows of 0.05 (bottom), 0.1 and 0.15 (top) W m^{-2}. For present day temperatures (210 K) melting occurs at a few hundred meters depths for local heat flows of 0.1 W m^{-2} or more.

[10] All models of Mars' thermal history show the surface heat flow rapidly declining at the end of heavy bombardment. If true, then heat flows significantly less than 0.1 W m^{-2} are likely for most of Mars' history after the end of heavy bombardment. For basal melting to occur during this period, kilometers thick accumulations of snow/ice would typically be needed [*Clifford*, 1987]. The scarcity of valley networks on Hesperian and Amazonian terrains is therefore understandable. Possible exceptions are the younger valley networks on several Hesperian/Amazonian volcanoes (Hecates, Ceraunius, Alba) [*Gulick*, 2001]. These may also have resulted from melting of ice deposits, perhaps triggered by local heat flows significantly higher than the global average.

4. Discussion and Conclusions

[11] Extensive equatorial snow and ice deposits may have been present in the late Noachian as a consequence of the large inventory of water on the surface and the cold global surface temperatures. As pointed out by *Clifford and Parker* [2001], high heat flows on early Mars would have resulted in much of the near-surface inventory being on the surface,

but as the heat flow declined with time, the surface inventory would have declined as progressively more water was frozen into the planet's cryosphere. Basal melting of these deposits may have provided the runoff needed to cut the valleys and recharge the groundwater system. Simulations indicate that even under present climatic conditions sub-aerial streams could flow for hundreds of kilometers under an ice cover provided certain minimum discharges are achieved [*Wallace and Sagan*, 1979; *Carr*, 1983]. Erosion by sub-snowpack streams, snowpack-fed sub-aerial streams, or spring-fed streams could all have contributed to formation of the valleys.

[12] Climate simulations and failure to detect weathering products on Mars suggest mostly sub-freezing conditions on early Mars. Higher heat flows and high near-surface water inventories on early Mars could have resulted in equatorial snow and ice deposits thick enough for basal melting to occur, thereby providing water to cut the valley networks.

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References

Acuna, M. H., et al., Global distribution of crustal magnetization discovered by the Mars Global Surveyor MAG/ER experiment, *Science*, *284*, 790–793, 1999.
Baker, V. R., Spring sapping and valley network development, *Geol. Soc. Sp. Paper*, *252*, 235–265, 1990.
Baker, V. R., Water and the martian landscape, *Nature*, *412*, 228–236, 2001.
Bandfield, J. L., T. D. Glotch, and P. R. Christensen, Spectroscopic identification of carbonate minerals in the martian dust, *Science*, *301*, 1084–1087, 2003.
Brailsford, A. D., and K. G. Major, The thermal conductivity of aggregates of various phases including porous materials, *Brit. J. Appl. Physics*, *15*, 313–319, 1964.
Carr, M. H., Stability of streams and lakes on Mars, *Icarus*, *56*, 476–495, 1983.
Carr, M. H., *Water on Mars*, Oxford, 229 p., 1996.
Carr, M. H., Elevations of water-worn features on Mars: Implications for circulation of groundwater, *J. Geophys. Res.*, *107*, 531, doi:10.1029/2002JE001845, 2002.
Choblet, G., and C. Sotin, Early transient cooling of Mars, *Geophys. Res. Lett.*, *28*, 3035–3038, 2001.
Christensen, P. R., et al., Mars Global Surveyor Thermal Emission Spectrometer experiment: Investigation description and surface science results, *J. Geophys. Res.*, *106*, 23,823–23,971, 2001.
Clifford, S. M., Polar basal melting on Mars, *J. Geophys. Res.*, *92*, 9135–9152, 1987.
Clifford, S. M., and T. J. Parker, The evolution of the martian hydrosphere: Implications for the gate of a primordial ocean and the current state of the northern plains, *Icarus*, *154*, 40–79, 2001.
Clow, G. D., Generation of liquid water on Mars through melting of a dusty snowpack, *Icarus*, *72*, 95–127, 1987.
Connerney, J. E. P., et al., Magnetic lineations in the ancient crust of Mars, *Science*, *284*, 794–798, 1999.
Costard, F., F. Forget, N. Mangold, and J. P. Peulvast, Formation of recent martian debris flows by melting of near-surface ground ice at high obliquity, *Science*, *295*, 110–113, 2002.
Craddock, R. A., and A. D. Howard, The case for rainfall on a warm, wet early Mars, *J. Geophys. Res.*, 107, 5111, doi:10.1029/2001JE0011505, 2002.
Gulick, V., Origin of valley networks on Mars: A hydrological perspective, *Geomorphology*, *37*, 241–268, 2001.
Haberle, R. M., Early Mars climate models, *J. Geophys. Res.*, *103*, 28,467–28,490, 1998.
Haberle, R. M., D. Tyler, C. P. McKay, and W. L. Davis, A model for the evolution of CO_2 on Mars, *Icarus*, *109*, 102–120, 1994.
Hynek, B. M., and R. J. Phillips, Evidence for extensive denudation of the Martian highlands, *Geology*, *29*, 407–410, 2001.
Jakosky, B. M., and M. H. Carr, Possible precipitation of ice at low latitudes of Mars during periods of high obliquity, *Nature*, *315*, 559–561, 1985.
Jakosky, B. M., and J. H. Jones, The history of martian volatiles, *Rev. Geophys*, *35*, 1–16, 1997.
Kasting, J. F., CO_2 condensation and the climate of early Mars, *Icarus*, *94*, 1–13, 1991.
McGovern, P. J., et al., Localized gravity/topography admittance and correlation spectra on Mars: Implications for regional and global evolution, *J. Geophys. Res.*, *107*, 5136, doi:10.1029/2002JE001854, 2002.
Mellor, M., Engineering properties of snow, *J. Glaciol.*, *19*, 15–66, 1977.
Melosh, H. J., and A. M. Vickery, Impact erosion of the primordial atmosphere of Mars, *Nature*, *338*, 487–489, 1989.
Mischna, M. A., M. I. Richardson, R. J. Wilson, and D. J. McCleese, On the orbital forcing of martian water and CO_2-cycles: A general circulation model study with simplified volatile schemes, *J. Geophys. Res.*, *108*(E6), 5062, doi:10.1029/2003JE002051, 2003.
Montesi, L., and M. Zuber, Clues to the lithospheric structure of Mars from wrinkle ridge sets and localization instability, *J. Geophys. Res.*, *108*, 5048, doi:10.1029/2002JE001974, 2003.
Newman, M. J., and R. T. Rood, Implications of solar evolution for Earth's early atmosphere, *Science*, *198*, 1035–1037, 1987.
Paterson, W. S. B., *The physics of glaciers* (3rd edition), Elsevier, 480 pp., 1994.
Pieri, D. C., Martian valleys: Morphology, distribution, age and origin, *Science*, *210*, 895–897, 1980.
Pitman, D., and B. Zuckerman, Effective thermal conductivity of snow at −88°, −25° and −5°C, *J. Appl. Physics*, *38*, 2698–2699, 1967.
Pollack, J. B., J. F. Kasting, S. M. Richardson, and K. Poliakoff, The case for a warm, wet climate on early Mars, *Icarus*, *71*, 203–224, 1987.
Presley, N. A., and P. R. Christensen, Thermal conductivity of particulate materials, 1. Review, *J. Geophys. Res.*, *102*, 6535–6549, 1997.
Schubert, G., S. C. Solomon, D. L. Turcotte, M. J. Drake, and N. H. Sleep, Origin and thermal evolution of Mars, in *Mars,* edited by H. H. Kieffer et al., Univ. of Arizona Press, 147–183, 1992.
Segura, T. L., O. B. Toon, A. Colaprete, and K. Zahnle, Environmental effects of large impacts, *Science*, *298*, 1977–1980, 2002.
Spohn, T., M. H. Acuna, D. Breuer, M. Golombek, R. Greeley, A. Halliday, E. Hauber, R. Jaumann, and F. Sohl, Geophysical constraints on the evolution of Mars, In *Chronology and Evolution of Mars*, edited by R. Kallenbach et al., Kluwer Academic Publ., Dordrecht, 231–262, 2001.
Wallace, D., and C. Sagan, Evaporation of ice in planetary atmospheres: Ice−covered rivers on Mars, *Icarus*, *39*, 385–400, 1979.
Zent, A. P., An open, snow-based hydrologic system as an analog for noachian Mars, *Lunar Planet. Sci.* [CD-ROM], *XXX*, abstract 1803, 1999.
Zuber, M. T., et al., Internal structure and early thermal evolution of Mars from Mars Global Surveyor topography and gravity, *Science*, *287*, 1788–1793, 2000.

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