Formation of Martian gullies by the action of liquid water flowing under current Martian environmental conditions

Jennifer L. Heldmann,[1,2] Owen B. Toon,[3] Wayne H. Pollard,[4] Michael T. Mellon,[1]
John Pitlick,[5] Christopher P. McKay,[6] and Dale T. Andersen[4]

Received 6 March 2004; revised 31 January 2005; accepted 9 March 2005; published 7 May 2005.

[1] Geomorphic evidence suggests that recent gullies on Mars were formed by fluvial activity. The Martian gully features are significant because their existence implies the presence of liquid water near the surface on Mars in geologically recent times. Irrespective of the ultimate source of the fluid carving the gullies, we seek to understand the behavior of this fluid after it reaches the Martian surface. We find that contrary to popular belief, the fluvially carved Martian gullies are consistent with formation conditions such as now occur on Mars, outside of the temperature-pressure stability regime of liquid water. Our model of the action of flowing pure liquid water produces the observed gully length distribution only at surface pressures and temperatures below the triple point where liquid water simultaneously boils and freezes and thus suggests that gullies were formed under conditions similar to present-day Mars. Our results suggest a typical flow rate of 30 m^3/s to carve the gully channels. At least 0.15 km^3 has flowed across the surface of Mars to carve the gully systems observed today, and this would require an aquifer recharge rate of $\sim 10^{-13}$–10^{-12} m/yr. The absence of gullies on Mars that are long enough to have been created above the triple point pressure argues that the atmospheric pressure has not been significantly larger than it is now since the origin of the gullies. This result may imply that Mars does not possess a significant reservoir of condensed CO_2.

Citation: Heldmann, J. L., O. B. Toon, W. H. Pollard, M. T. Mellon, J. Pitlick, C. P. McKay, and D. T. Andersen (2005), Formation of Martian gullies by the action of liquid water flowing under current Martian environmental conditions, *J. Geophys. Res.*, *110*, E05004, doi:10.1029/2004JE002261.

1. Introduction

[2] Images from the Mars Orbiter Camera (MOC) on the Mars Global Surveyor (MGS) spacecraft show geologically young small-scale features resembling terrestrial water-carved gullies (Figure 1a). An improved understanding of these features has the potential to reveal important information about the hydrological system on Mars, which is of general interest to the planetary science community as well as the field of astrobiology and the search for life on Mars. The young geologic age of these gullies is often thought to be a paradox because liquid water is unstable at the Martian surface. Gullies are generally found where surface temperatures and pressures are below the triple point of water (273 K, 6.1 mbar) so that liquid water will spontaneously boil and/or freeze [*Haberle et al.*, 2001]. Furthermore, gullies are generally found where temperature and pressure conditions on the Martian surface never exceed the triple point at any time during the year [*Haberle et al.*, 2001]. We therefore examine the flow of water on Mars to determine what conditions are consistent with the observed features of the gullies. Previous researchers proposed numerous mechanisms to supply water (and other fluids) to form the gullies [*Malin and Edgett*, 2000; *Gaidos*, 2001; *Mellon and Phillips*, 2001; *Musselwhite et al.*, 2001; *Bridges and Hecht*, 2002; *Costard et al.*, 2002; *Gilmore and Phillips*, 2002; *Knauth and Burt*, 2002; *Heldmann and Mellon*, 2004; *Treiman*, 2003; *Christensen*, 2003; *Hecht*, 2002] and so the supply of the water is an issue we do not address here. Some of these researchers have assumed that the gully features formed under different climatic conditions when liquid water was stabilized. A considerable amount of work also has been done to investigate large-scale fluvial features on Mars, some of which extend 1000 km. Researchers have shown that brines or ices that cap relatively slowly flowing streams can stabilize liquid water under current conditions which can explain the valley network formation [*Knauth and Burt*, 2002; *Wallace and Sagan*, 1979; *Carr*, 1983]. In contrast, we find that the short length of the gully features implies they did form under conditions similar to those on present-day Mars by simultaneous

[1]Laboratory for Atmospheric and Space Physics, University of Colorado, Boulder, Colorado, USA.
[2]Now at Space Sciences Division, NASA Ames Research Center, Moffett Field, California, USA.
[3]Program in Atmospheric and Oceanic Sciences, Laboratory for Atmospheric and Space Physics, University of Colorado, Boulder, Colorado, USA.
[4]Department of Geography, McGill University, Montreal, Quebec, Canada.
[5]Department of Geography, University of Colorado, Boulder, Colorado, USA.
[6]Space Sciences Division, NASA Ames Research Center, Moffett Field, California, USA.

0148-0227/05/2004JE002261$09.00

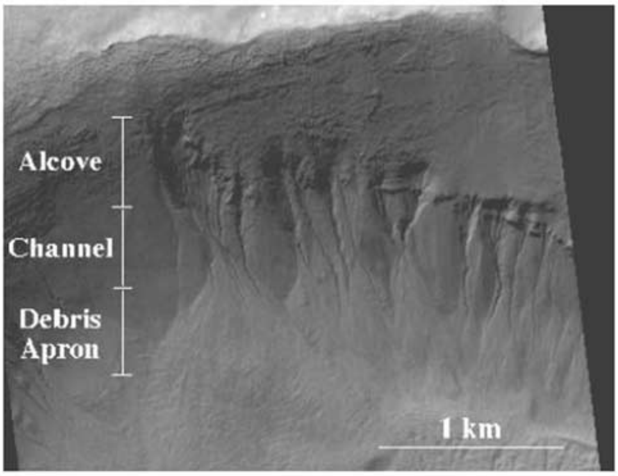

Figure 1a. Portion of MOC image M17-00423 located at 200.86°W, 39.16°S showing the alcove, channel, and debris apron structures of recent gullies on Mars. Scale bar is 1 km.

freezing and rapid evaporation of nearly pure liquid water.

2. Martian Gullies

[3] Narrow-angle MOC images show gully features in the mid to high latitudes (30°–72°) of both the northern and southern hemispheres [*Malin and Edgett*, 2000; *Heldmann and Mellon*, 2004; *Haberle et al.*, 2001]. The characteristic alcove, channel, and debris apron morphologies suggest liquid water is the erosional agent while the dearth of overlying impact craters and superposition of gully features upon young sand dunes and polygonal ground suggest the geologically recent age of the gullies. The gully features generally lie at the average elevation for their given latitude and are not found at the extreme high or low elevations [*Heldmann and Mellon*, 2004]. Liquid water is typically stable at the lowest elevations and at low latitudes on the planet because the atmospheric pressure is greater than the vapor pressure of water and surface temperatures in equatorial regions can reach 273 K for parts of the day [*Haberle et al.*, 2001]. However, gullies are not found in these low elevation regions. MOLA elevations were extracted from 1/20° resolution binned maps for all 106 MOC images containing gullies used in this study. Gullies generally tend to form outside of the pressure stability field for liquid water as 70% of the gully systems examined lie above the triple point elevation on Mars. Of the 30% of the gully systems currently below this elevation, 16% of these low elevation gullies lie within 1000 m of this datum. These low-elevation gullies are almost exclusively within Dao Vallis, a large outflow channel beginning near the Hadriaca Patera volcanic complex and extending into the edge of the Hellas Basin. Only two of these low elevation gullies are not within Dao Vallis and they are located within a crater north of Reull Vallis and within another valley system west of Hadriaca Patera.

[4] There are several types of gullies on Mars. Those on sand dunes, not considered here, have large leveed mounds along their flanks suggesting a fluidized, or other relatively dry flow. The gullies we consider typically have the classic eroded alcove, incised channel, and debris apron structure, suggesting the flow was dominated by fluid rather than debris [*Malin and Edgett*, 2000; *Heldmann and Mellon*, 2004] (Figure 1a). The classic interpretation of the geomorphic alcove-channel-debris apron structure is that the fluid emanates from within the alcove causing headward erosion to create the alcove structure. The fluid flows downslope to carve the channels and any debris that may be mobilized is deposited in the debris apron (see summary by *Heldmann and Mellon* [2004]).

[5] There are several interesting features of the gullies which were discussed by *Heldmann and Mellon* [2004] that are critical to our understanding of gully formation. First, the gully alcove bases are located at a depth below the overlying ridge that corresponds with the freezing isotherm in the Martian regolith. On the basis of the measured TES thermal inertia (I) at each gully site as well as modeled mean surface temperatures (T_o) [*Mellon et al.*, 2000], *Heldmann and Mellon* [2004] computed the subsurface temperature at the depth of each individual alcove base. For a given alcove base depth (z) the aquifer temperature is calculated from conduction of geothermal heat as $T = (q/k)z + T_o$ using values for geothermal heat flux (q) and thermal conductivity (k) from Table 1. These thermal conductivities were calculated on the basis of the TES thermal inertia at each location since thermal inertia (I), thermal conductivity (k), density (ρ), and specific heat (c) are related by $I = \sqrt{k\rho c}$. The correlation of the gully alcove base depths and the computed depth to the freezing isotherm implies the gullies tap into an underground reservoir of liquid water on Mars [*Heldmann and Mellon*, 2004]. Second, the gully channels are remarkably short, averaging 500 m in length [*Heldmann and Mellon*, 2004]. In this work we have also reexamined the *Heldmann and Mellon* [2004] data set to identify a third interesting feature. We have analyzed the images studied by *Heldmann and Mellon* [2004] with adequate MOC and MOLA (Mars Orbiter Laser Altimeter) data coverage and find that the channels in 80% of the gully systems do not reach the end of the slope and hence channel length is not limited by topographic control. For example, Figure 1b shows MOC wide- and narrow-angle images where gullies are found on the south facing wall of Dao Vallis. The distal end of the gullies occurs before the bottom of the slope. It is the gully channel length that we wish to model here in order to constrain the properties of the flowing liquid.

Table 1. Subsurface Temperature Calculations: Model Parameters

Property	Parameter	Dry Soil	Units
Soil density	ρ	1650[a]	kg/m^3
Heat capacity	c	837[a]	J/kg-K
Thermal conductivity	k	$I^2/\rho c$	W/m-K
Geothermal heat flux[b]	q	30	mW/m^2

[a]*Mellon and Phillips* [2001].

[b]The Martian geothermal heat flux (q) has never been measured, but theoretical values range from 20 to 45 mW/m^2, and so we adopt the typical value of 30 mW/m^2 [*Johnston et al.*, 1974; *Fanale*, 1976; *Toksoz and Hsui*, 1978; *Toksoz et al.*, 1978; *Davies and Arvidson*, 1981; *Stevenson et al.*, 1983; *Franck and Orgzall* 1987; *Schubert and Spohn*, 1990; *Spohn*, 1991]. Results are not very sensitive to the value of q. See *Mellon and Phillips* [2001] for further discussion of geothermal heat values.

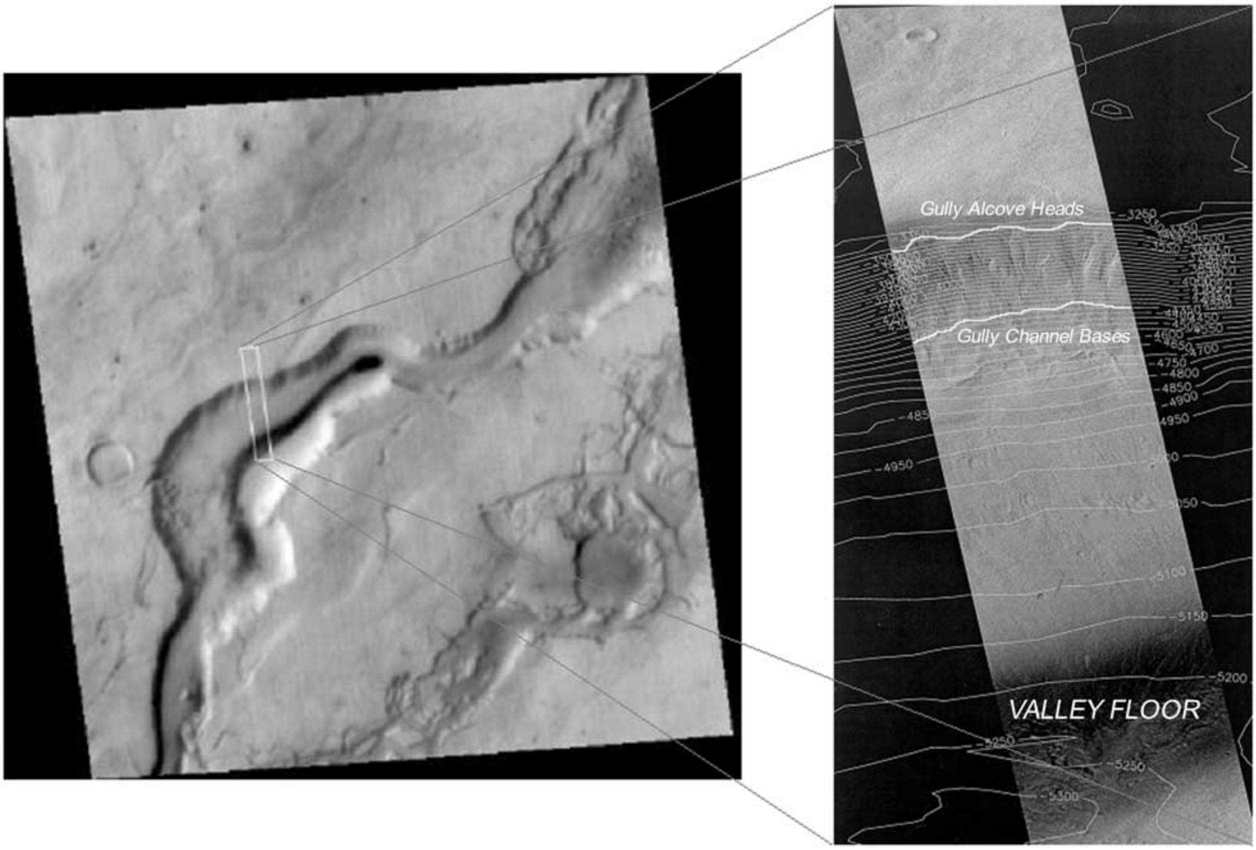

Figure 1b. MOC wide-angle (WA) context image M07-03140 and MOC narrow-angle (NA) image M07-03139 (NA image center is at 269.91°W, 35.89°S; image width is 2.85 km). MOLA contour lines are superposed on the MOC NA image to show elevations in meters. Gullies are located on the south facing wall of Dao Vallis. Gully alcove heads are shown by the top white line on the NA image. Gully channel bases are shown by the bottom white line on the NA image. Note the gully channels do not extend to the base of the slope on the valley floor.

[6] As discussed in the literature, the liquid most likely originates in the alcove, flows downslope through the channel, and then deposits debris in a distal debris apron [*Malin and Edgett*, 2000; *Mellon and Phillips*, 2001; *Gilmore and Phillips*, 2002; *Heldmann and Mellon*, 2004]. If the base of the channel were where the channel simply fills in with sediment, then one would expect to see evidence of pooling of water above the debris field. Instead gully channels are clearly eroded into the host rock and taper downslope. The channel therefore ends where not enough liquid is available to continue eroding into the preexisting rock. Some channels do not end with debris aprons and so clearly they are not being choked with debris. Instead the channels taper to an end and are no longer incised into the host rock, implying that there is no more fluid remaining to cause the erosion. Channels that do have distal debris aprons still taper to an end implying the loss of fluid due to freezing and evaporative processes. If the fluid were not leaving the channel then this liquid would still have to go somewhere. Given the dimensions of the channels (10 m wide x 10 m deep) there should be a downslope manifestation of this water flow which is not observed. Instead the water disappears from the system; our interpretation of these observations is consistent with our model results.

3. Model

[7] We numerically simulate the flow of liquid water within a Martian gully channel. Our goal is to determine whether liquid water can flow over sufficient distances to carve the observed channels and also to place constraints on the flow rate and salinity of the water. This model is first developed to simulate a well-observed terrestrial example of channel flow in the High Canadian Arctic. This environment is an analog to the one commonly assumed by other researchers who have suggested the Martian flow occurred in a different climatic regime with higher pressures so that the liquid did not rapidly evaporate during its flow. Our nominal Mars simulation assumes an atmospheric pressure below the triple point pressure which results in a rapid evaporation of liquid water. For completeness we also consider the case where the water does not rapidly boil (due to higher atmospheric pressure and/or the presence of salts).

[8] We developed a one dimensional model of channel flow by dividing the channel into equally spaced parcels

and performed an energy balance on each individual parcel. The flow is modeled as a Newtonian fluid whose downslope velocity is determined by a balance between the weight of water in each parcel and the turbulent shearing resistance. Changes in channel shape or bed elevation due to erosion and sediment transport are ignored, although these processes may be important in forming the gullies initially. Energy source terms calculated for each parcel include solar insolation, radiative and sensible heating by the atmosphere, and the latent heat of fusion based on ice formation. Energy loss terms include the latent heat of vaporization as water evaporates, radiative cooling of the liquid, and thermal conduction from the liquid to the ground.

[9] Liquid water is lost from the channel by both freezing and evaporation. Water lost from the channel via infiltration to the underlying soil is negligible for the Arctic and assumed negligible for the Martian cases due to the slope angle of the channel. Using the data set of *Heldmann and Mellon* [2004], average gully channel slopes were determined by dividing the vertical elevation change of the channel by the horizontal channel length. Average channel slopes on Mars are ~18° with a standard deviation of 10°. Laboratory studies indicate that infiltration rates on standard soil at slopes of 16° are typically less than 100 mm/hr and decrease with increasing slope angle [*Dingman*, 1994]. Channel slopes do not change substantially along the length of the channel and given the resolution of the MOLA data (160 m footprint) and MOC resolution (best values of 1–2 m per pixel) the channels maintain a relatively flat profile.

[10] For cases where the atmospheric pressure is greater than the vapor pressure of water (i.e., terrestrial cases and Mars cases involving brines), evaporation rates are calculated using a bulk aerodynamic term for vertical kinematic turbulent water vapor flux. The water vapor mass flux due to evaporation (E_1) is thus calculated using

$$E_1 = C_e\, v_h(r)|[\rho_v(z_o) - \rho_v(z_r)], \tag{1}$$

where C_e is the bulk transfer coefficient of water vapor equal to 0.015, v_h is the horizontal wind velocity (assumed to be 1 m/s), and ρ_v is the total atmospheric density at the surface (z_o) and at a reference height (z_r) of typically 10 m [*Jacobson*, 1999].

[11] For cases where the atmospheric pressure is less than the vapor pressure of the water (i.e., Martian cases involving pure water), the liquid water loss rate is calculated using kinetic theory, assuming evaporation under vacuum [*Wallace and Sagan*, 1979; *Kennard*, 1938]. The resulting evaporation mass flux (E_2) is given by

$$E_2 = \alpha V_s[\rho_v(z_o) - \rho_v(z_r)], \tag{2}$$

where V_s is the mean molecular speed of the molecule, ρ_v is the water vapor density (assumed to be zero at the reference height), and α is an empirical coefficient which has a value near unity [*Tschudin*, 1946]. The value of α (the coefficient of evaporation) has been determined empirically and is 0.94 ± 0.06. We therefore adopt $\alpha = 1$, which is valid when $p_v > p_a$, which is true on present-day Mars where ambient surface pressures are below the triple point pressure of water. As the liquid water evaporates, the water vapor has a higher pressure than the surrounding gas that will create a hydrodynamic flow in the gas. In our everyday experience with boiling the fluid temperature is raised until the vapor pressure just slightly exceeds that of the gas so the fluid evaporation rate is relatively gentle. In the Martian case fluid well above the boiling point is suddenly exposed to the atmosphere. The difference between the vapor and ambient pressures relative to the ambient pressure is large and flash boiling occurs, leading to a violent loss of fluid. It is possible that even for pressures slightly above the triple point that the water will rapidly evaporate. However, it is not our goal here to explore the threshold near the boiling point to determine just how close one can be and still have the channel length limited by boiling and freezing. New experimental data may be needed for such studies. Also, if mixing of water vapor into the atmosphere is limited due to the turbulent nature of the flowing liquid or some other unconsidered effect(s) then rates of freezing and evaporation may be different than suggested by our model and could alter the predicted flow rates accordingly.

[12] The drop in temperature of the liquid water caused by the evaporation is calculated from the mass evaporation flux and the latent heat of vaporization. The ice that forms from the freezing of the brine within each parcel is observed to accumulate at the edges of the channel in the Arctic case due to the relatively low flow velocities and hence forms an ice cover which inhibits further evaporation [*Heldmann et al.*, 2005]. For the Mars case the high flow velocities due to the relatively steep channel slopes as well as vigorous evaporation of the water in some cases does not allow for ice accumulation within the channels and so the ice that forms within the model is nominally considered lost from the system and does not form an ice cap over the channel. However, we do also consider the case where ice remains within the channel and is transported downslope to quantify the effects of an ice-laden flow. Variable input parameters for both the terrestrial and Martian cases related to the flow itself include flow rate, channel volume, initial outlet temperature, and initial salinity of the liquid.

4. Results for Earth

[13] We tested our model using known flow parameters and environmental conditions of perennial saline springs in the Mars analog environment of the Canadian High Arctic. The springs at Gypsum Hill on Axel Heiberg Island are located at 79°24′30″N, 90°43′05″W in a polar desert environment where the mean annual air temperature is 258 K and winter temperatures of 218 K are common [*Doran et al.*, 1996; *Andersen et al.*, 2002; *Maxwell*, 1982]. This High Arctic site is classified as a polar desert since evaporation exceeds the low annual precipitation [*Andersen et al.*, 2002]. The Gypsum Hill springs flow continuously throughout the year [*Heldmann et al.*, 2005]. A typical channel has a width of 0.2 m and the liquid water flows 600–625 m [*Heldmann et al.*, 2005]. The brine solution has an observed flow rate of $\sim 1 \times 10^{-3}$ m^3/s, initial outlet temperature of 278 K, an initial salt mole fraction of 0.06, and an average flow velocity of ~1 m/s [*Heldmann et al.*, 2005; *Pollard et al.*, 1999].

[14] Numerical simulations for this standard Arctic case indicate that liquid brine simultaneously evaporates and freezes which increases the salinity of the remaining liquid

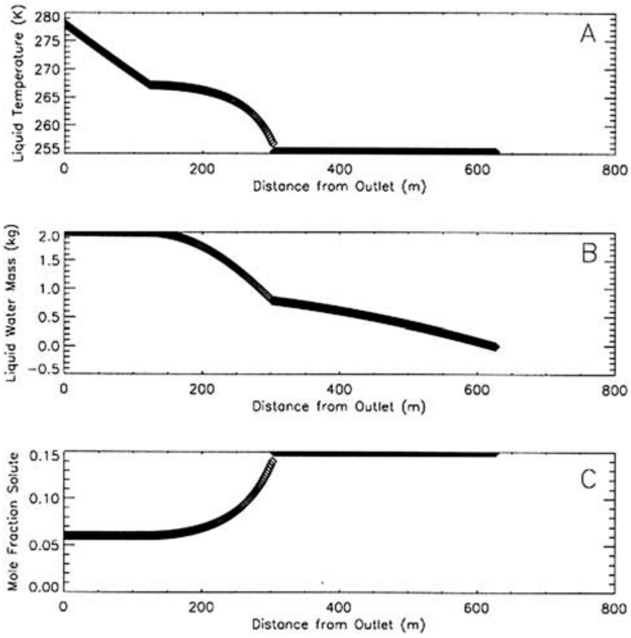

Figure 2. The (a) liquid water temperature, (b) liquid water mass, and (c) brine mole fraction of solute as a function of distance from the channel outlet for the standard Arctic case.

within the channel as shown in Figure 2. In this model the behavior of the liquid brine is divided into three phases, which include (1) cooling down to the freezing point, (2) the mutual coexistence of ice and liquid below the freezing point but above the eutectic point, and (3) the precipitation of salt from the solution at the eutectic point. At the end of Phase 1, the water cools down to the freezing point of the solution which is dependent on the mole fraction of solute present in the liquid. During this phase, liquid water is lost only via evaporation. Evaporative losses are negligible as evidenced by the low mass loss from 0–175 m down the length of the channel as shown in Figure 2b. Once the liquid reaches the freezing point, ice begins to form in the channel. More liquid water mass is lost once the solution reaches the freezing point and ice begins to form as shown by the dramatic drop in liquid water mass further along the length of the channel (Figure 2b). The amount of ice formed is governed by an energy balance among evaporation, cooling, and ice formation. The amount of ice formed also depends on the relationship between temperature and mole fraction of solute through the freezing point depression curve. The amount of ice formed is determined by a convergence of the energy balance and freezing point depression equations. As ice forms, the remaining liquid solution becomes more concentrated and hence ice forms at increasingly lower temperatures due to the increased freezing point depression. This process continues until the solution reaches the eutectic point where salt begins to precipitate out of the solution as ice forms and the salinity remains constant at this maximum value.

[15] The resulting modeled channel for this standard Arctic case spans 624 m in length before all the liquid is lost to evaporation and ice formation. The maximum distance the liquid brine flows at the Gypsum Hill springs during the winter months is 600–625 m [*Heldmann et al.*, 2005] and so our calculated runout value is in excellent agreement with the physical dimensions of the Arctic spring system.

[16] Several alternative simulations were run to determine the effects on channel length of variable input parameters including incoming insolation, ambient atmospheric and ground temperature (assumed equal), initial liquid outlet temperature, and liquid velocity. Each parameter is varied independently and compared with the standard Arctic case previously described.

[17] Realistic variations in insolation and initial outlet temperature have little effect on the runout distance of an Arctic channel. Figure 3a shows that insolations (solar and downwelling infrared) ranging from 50–300 W/m^2 cause the total channel runout distance to vary by only 40 m. Figure 3b shows that variations in channel length due to changes in the initial liquid temperature which could be caused by subsurface aquifer conditions are also minimal.

[18] Variations in atmospheric temperature, ground temperature, and flow velocity have significant effects on channel runout distances as shown in Figures 3c and 3d. However, direct measurements of ambient temperatures in the Arctic have been made and this parameter is well known with average air temperatures in late April of ~243 K [*Heldmann et al.*, 2005]. Flow velocity of the Arctic spring is also well constrained at ~1 m/s based on direct observation.

5. Results for Mars

[19] We now apply our model to various cases on Mars. To place constraints on Martian channel dimensions, we analyzed 106 MOC narrow-angle images containing gullies located in the southern hemisphere [*Kirk et al.*, 2001]. The MOC data were radiometrically calibrated and geometrically transformed to an equal-area sinusoidal map projection using ISIS (Integrated Software for Images and Spectrometers) [*Kirk et al.*, 2001]. Linear distances were determined directly from the ISIS-projected MOC image. Channel lengths were measured from the alcove base to the beginning of the debris apron or until the channel ended if there was no clear debris apron [*Heldmann and Mellon*, 2004]. Typical gully channel widths on Mars are 10 m. Of the 276 Martian gully channels measured, 250 of these channels have lengths less than 1 km. The longer channels are only found equatorward of 55°S but otherwise have the same characteristics as the shorter channels. The average channel length is 500 m. The longest channel length is 3.15 km. These measured channel lengths are indicative of the distance of liquid water flow on the Martian surface.

[20] Flow velocities are calculated on the basis of the slope of the channels derived from combined MGS MOC and Mars Orbiter Laser Altimeter (MOLA) data. Channel slope angles were determined by dividing the vertical elevation change of the alcove by the horizontal channel length [*Heldmann and Mellon*, 2004]. From both the vertical and horizontal dimensions of the gullies extracted from the MGS data, the average channel slope is found to be ~18°.

[21] The mean velocity in a turbulent, clear-water (i.e., low sediment load) flow is given by *Dingman* [1984] as v =

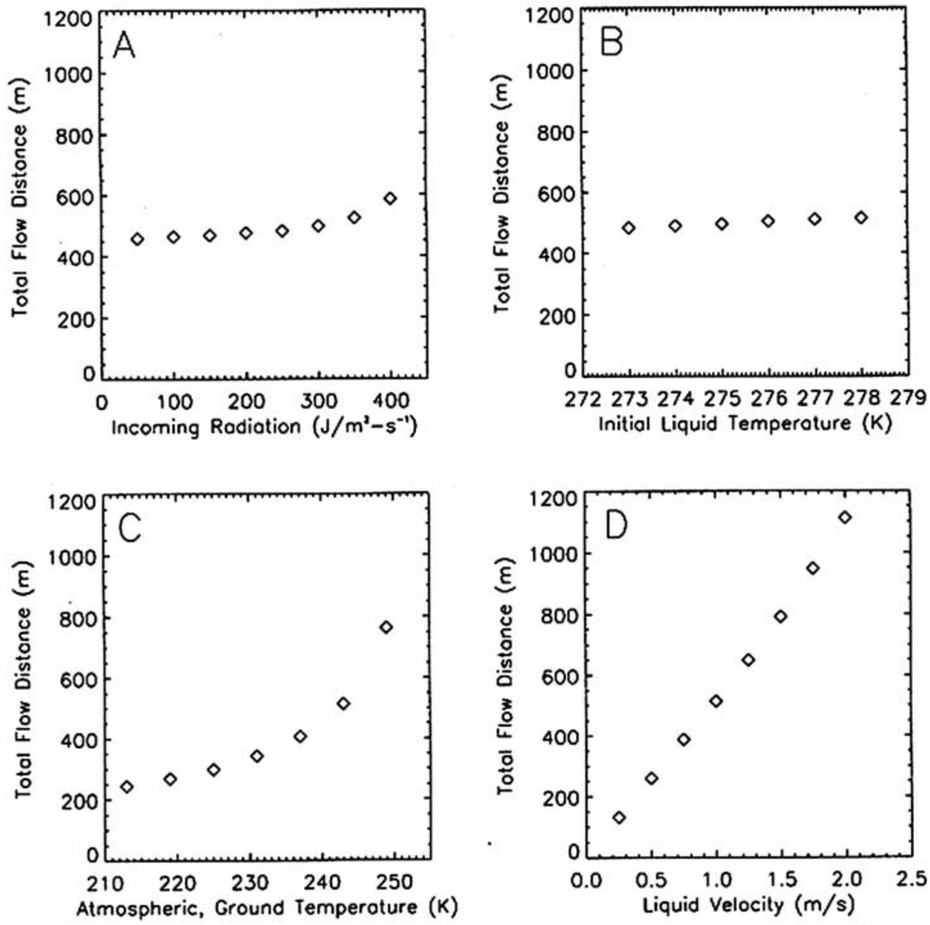

Figure 3. The effects of varying (a) insolation, (b) initial outlet temperature, (c) atmospheric and ground temperature, and (d) liquid velocity on the final liquid runout distance in each numerical simulation.

$2.5V^*\ln(Y/k_s) + 6.0$, where k_s is the channel bed roughness and V^* is the friction velocity calculated as $(gYS)^{1/2}$ where g is the gravitational acceleration, Y is the depth, and S is the slope of the channel. Using average channel slope values (~18°) derived from the MOC and MOLA data sets, channel bed roughness values on the order of 1 m (observed channels are relatively smooth at the highest resolution (~1.5 m/pixel) of MOC imaging), as well as plausible water depths yields flow velocities on the order of 10 m/s.

[22] In our simulations, a range of initial flow rates is used to represent channel depths ranging from 0.15 m–1 m. A range of initial salinities from pure water to highly saline brines similar to the Arctic brines (~5 times the salinity of seawater) is also used to place constraints on the concentration of a possible brine.

[23] Liquid water is unstable due to the low ambient pressure and temperature conditions at most locations on Mars including most locations of gully occurrence [*Heldmann and Mellon*, 2004; *Haberle et al.*, 2001]. The presence of soluble salts in a solution, however, results in vapor pressure lowering of the solution. A salt solution at an atmospheric pressure of 6 mbar requires 0.02 mole fraction of solute (approximately twice the salinity of terrestrial seawater) to maintain a vapor pressure comparable with the ambient atmospheric pressure on the Martian surface at 273 K.

[24] Numerical simulations conducted for the standard Mars case using an initial outlet salinity of 0.02 with a conservatively low flow rate corresponding to a channel brine depth of 0.5 m result in channel runout distances spanning tens of kilometers. Liquid water is lost mainly by the formation of ice as the channel freezes. Such long runout distances are consistent with early modeling efforts of *Wallace and Sagan* [1979] and *Carr* [1983] to explain the formation of the valley network systems. In their models pure liquid water is inhibited from rapidly escaping to the atmosphere due to the formation of a protective ice cover whereas in this work such evaporation rates into the atmosphere are decreased by the presence of soluble salts. In each case, however, the end result is the same; liquid water can run for extensive distances (i.e., tens of kilometers) on the Martian surface if this evaporation is suppressed.

[25] We now consider the case for pure liquid water on Mars where evaporation is allowed to rigorously occur. Results from a simulation of channel flow on Mars using a flow rate of 30 m^3/s (0.3 m water depth in a channel 10 m wide with a flow velocity of 10 m/s) for pure water are shown in Figure 4. In this case the liquid water flows for a distance of 546 m which is approximately the average length of a gully channel on Mars. Non-boiling evaporative losses on Mars are extremely small with values of 1.0 ×

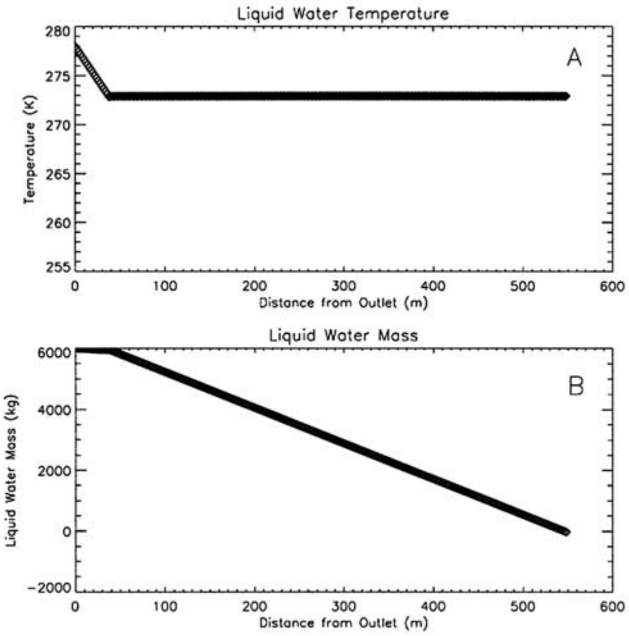

Figure 4. The (a) liquid water temperature and (b) liquid water mass as a function of distance from the channel outlet for the Mars case involving pure water.

10^{-6} kg/s-m^2. Some liquid water is lost from the channel by considering evaporation into a vacuum as 0.68 kg/s-m^2 escapes to the atmosphere. The majority of the liquid is lost to ice formation at a rate of 5.15 kg/s-m^2.

[26] Simulations are run for a variety of flow rates for the pure water case. As shown in Figure 5, flow rates ranging from 15 m^3/s–60 m^3/s (0.15 m–0.6 m water depth) result in channel runout lengths of 272 m–1092 m. These distances are consistent with the channel lengths of observed Martian gullies. Using the measured length of individual channels, the minimum flow rate of individual gullies can thus be determined.

[27] Our calculated flow rates are consistent with flow estimates based on the geomorphology of the Martian gullies. On the basis of the size of the debris aprons and the corresponding amount of water required to create these features, *Malin and Edgett* [2000] estimate that 2500 m^3 of water reached the end of each channel. For the standard Martian gully channel with a length of ~500 m, Figure 5 indicates a flow rate of 30 m^3/s. Given a flow velocity of 10 m/s, water will reach the end of the channel in 50 seconds. Assuming a continuous source of flow for these 50 seconds yields a total flow volume of 1500 m^3 which is comparable with the flow volume previously estimated by *Malin and Edgett* [2000]. However, actual durations of flow are most likely 7–10 times greater than this estimate because of the large volumes of ice formed down the length of the channel. The amount of ice formed in the channel dominates compared to liquid lost to evaporation. The amount of ice formed is primarily determined by the ratio between the latent heats of fusion and vaporization as freezing is balanced by evaporation. Since the ice does not flow, the minimum flow time is increased by the ratio of ice to liquid and results in a factor of 7–10 increase in flow duration to be consistent with the geomorphology of the gullies. The debris aprons are most likely formed by the mass movement of material entrained in liquid water and hence water probably flowed through the channels for a minimum of 500 seconds (8.3 minutes) to form the aprons. Assuming the water initially flowing through the channels consists of 10% sediment [*Malin and Edgett*, 2000] then under the nominal flow rate of 30 m^3/s the remaining water becomes laden with 50% debris 494 m down the length of the channel. Sediment can therefore be carried through the channel and deposited in the debris apron ~500 m from the alcove source region as observed in MOC imagery. The results of our modeling efforts are largely consistent with the geomorphology of the gully systems and suggest that the gullies formed from short outbursts which episodically released water from a subsurface aquifer as opposed to longer duration flows.

[28] We now consider the case where the ice crystals remain within the system and are carried downstream within the flow. This ice that remains in the channel eventually chokes the liquid water flow. In this case the atmospheric pressure is still less than the vapor pressure of the water which results in the rapid evaporation rates. The channel would consist of an equal portion of ice and liquid 220 m down the length of the channel given the nominal flow rate of 30 m^3/s. However, increasing the flow rate to 60 m^3/s results in a channel runout distance of 440 m before the amount of ice present is comparable to the amount of liquid water left in the channel. The presence of 50% ice would begin to choke the liquid flow and thus would increase the required outlet flow rate by a factor of two to obtain the same channel length. This finding further supports the notion that the liquid water must be simultaneously freezing and evaporating under the low pressure and low temperature conditions of Mars to create the relatively short gully channels because otherwise such flow rates result in channels that are

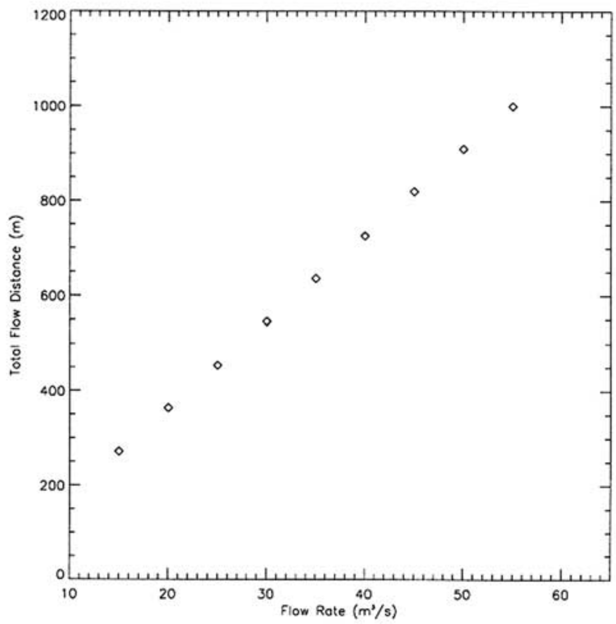

Figure 5. Flow rate versus channel runout distance for the Mars case.

tens of kilometers long before they become clogged with ice. Additional factors such as infiltration rates, stronger atmospheric mixing, and/or weaker brine solutions could also accelerate liquid water loss and hence account for shorter channel lengths.

[29] Simulations of channel flow on Mars where rapid evaporative losses are suppressed indicate that the flow rate necessary to create a channel 500 m in length is 0.005 m^3/s (corresponding to a channel depth of 50 microns). Such low flow rates are not likely to have carved the relatively large (10 m diameter) channels in a short period of geologic time. It is also not geologically possible that such a trickle of water could create the large (several hundred meters in length and width [*Heldmann and Mellon*, 2004]) eroded alcoves observed at the heads of the gully channels. The scenario of a salty brine carving the Martian channels, or channel formation under a higher pressure Martian climate, is unrealistic. Likewise, water flowing under a protective sheet of snow and/or ice which would inhibit evaporation into the atmosphere is similarly unrealistic. The model results are suggestive of an open channel of liquid water flow exposed to the ambient Martian environmental conditions. Of course if the ice is limited in extent the channel would soon disappear once it emerged from the ice sheet.

[30] On the basis of the total volume of water flowing through each gully and the total number of gullies discovered on Mars, we can estimate the minimum amount of water that must have been sequestered in subsurface aquifers and an aquifer recharge rate. Over 10,000 individual gullies have been observed on Mars [*Edgett et al.*, 2003]. Assuming a typical flow rate of 30 m^3/s for 500 seconds per gully then 0.15 km^3 has flowed across the surface of Mars to carve the gully systems observed today. This is enough water to cover the south polar cap on Mars to a depth of ~1 mm. If we assume that the typical age of the gullies is 10^6–10^7 years and this represents their lifetimes then about 10 to 100 km^3 of water has moved through the gullies over geologic time. In order to have gullies still active on Mars, the aquifer must be recharged at a rate of 10^1–10^2 m^3/yr. Assuming the recharge occurs by vapor diffusion over the entire planet, a layer of water $\sim 10^{-13}$–10^{-12} m/yr must reach the subsurface which is a very small fraction of the annual atmospheric column of water of 10 microns. Hence aquifer recharge to explain the gullies is not a significant problem and may easily occur via vapor diffusion into the soil. The total inventory of water on Mars attributed to gully formation is most likely higher than we estimate and can be increased by higher flow rates due to evaporation and longer durations of flow needed because of ice formation.

6. Conclusions

[31] Geomorphic evidence suggests that recent gullies on Mars were formed by fluvial activity. Likewise, the depth to the gully alcove regions is remarkably consistent with the depth to the 273 K isotherm within the Martian subsurface and hence liquid water could be present in a subsurface aquifer at these locations [*Heldmann and Mellon*, 2004]. Numerical simulations show that pure liquid water flowing at rates of 15–60 m^3/s is consistent with these observations. The formation of gullies on Mars is inconsistent with briny fluid flows with significant flow rates because inhibiting rapid evaporation by vapor pressure suppression (or other means such as ice sheets capping the flow, or a higher pressure climate state) results in channels that are much longer than those observed on Mars. Instead, our model indicates that these fluvially carved gullies formed in the low temperature and low pressure conditions of present-day Mars by the action of relatively pure liquid water.

[32] These model results suggest that the gullies formed where conditions are above the triple point elevation to enable the rapid evaporative loss of the liquid water on the Martian surface. Gullies located below the triple point altitude are located primarily in Dao Vallis and have no other outstanding characteristics (including channel length) with respect to gullies found at higher elevations [*Heldmann and Mellon*, 2004]. According to recent Mars climate models, the surface pressure of Mars fluctuates significantly on 120 kyr periods due to obliquity oscillations [*Laskar et al.*, 2002]. Therefore the triple pressure altitude may fluctuate and so perhaps these lower elevation gullies formed at a different time than their higher elevation counterparts. This scenario would remove an altitude dependence on morphology and is consistent with the gully elevations and channel lengths observed today.

[33] A surprising result, in light of the conclusions in this paper, is that *Heldmann and Mellon* [2004] do not see a set of long gullies which correspond to flow during times of high obliquity when pressures may have been significantly higher on Mars. We suggest three possible reasons for the lack of such longer gullies. First, it might be that the formation mechanism for the gullies only operates at low obliquities when atmospheric pressures are low. Second, the gullies may be much younger than has been thought. The last time the obliquity was large on Mars was at least 250,000 years ago. It may be that the long gullies which formed at that time have simply been eroded away or buried by dust since then. Finally, it may be that there is too little CO_2 in the Martian polar caps for the pressure to be much higher than it is now.

[34] Of these three possibilities the third is most interesting because it would suggest that much of our current thinking about the past history of Mars climate is incorrect. This possibility is also the most plausible. *Byrne and Ingersoll* [2002] have argued that only about 0.36 mbar of CO_2 is present in the residual polar caps on Mars, on the basis of modeling pits in the South Polar cap. If so, the Martian pressure in the past cannot have been much greater than it is now. Our gully model is consistent with that result.

[35] Some gullies now exist on Mars at elevations where pressures may be slightly above the triple point of water for parts of the Martian year. We expect that rapid boiling is not a step function process that occurs only at the triple point, but probably also occurs for pressures slightly above it. Future laboratory and theoretical work is needed to clarify the pressure and temperature regimes for rapid evaporative loss. Alternatively the lower elevation gullies may also have formed at a time of year, or part of the obliquity cycle in which the pressure was lower than the current average value. The atmospheric pressure on Mars at any location varies seasonally [*Haberle et al.*, 2001]. Therefore even gullies forming below the triple point elevation on Mars may actually form at times of the year when the atmospheric pressure is relatively low (lower than the triple point

pressure) and therefore the liquid water would experience rapid boiling. If gullies do form at atmospheric pressures greater than the triple point pressure then our model does not explain these low altitude features.

[36] The other two possibilities seem less likely. No current theory of gully formation predicts they would form preferentially at low obliquity. Perhaps the lower pressure reduces the overload and makes springs more likely to form, but the mass of the atmosphere on Mars is only equivalent to about 5–10 cm of Martian soil so it is not much of a load on the soil. The current Martian erosion rates and dust deposition rates are not sufficient to bury or erase gullies thought to be 10 m deep in only 250,000 years [*Golombek and Bridges*, 2000].

[37] Both the TES (Thermal Emission Spectrometer aboard the Mars Global Surveyor spacecraft) and THEMIS (Thermal Emission Imaging System aboard the Mars Odyssey spacecraft) instruments have failed to detect salt deposits associated with the gullies. This further supports our findings of relatively pure water flow on Mars although the spatial and spectral resolution of these instruments is not optimal for detection of salt deposits associated with the gully features.

[38] The gullies may be sites of near-surface water on present-day Mars and should be considered as prime astrobiological target sites for future exploration. Typical gully channels lie on modest slopes of ~18° and future robotic rovers could traverse such terrain to search for evidence of past or present life on Mars.

References

Andersen, D. T., W. H. Pollard, C. P. McKay, and J. Heldmann (2002), Cold springs in permafrost on Earth and Mars, *J. Geophys. Res.*, *107*(E3), 5015, doi:10.1029/2000JE001436.

Bridges, N. T., and M. H. Hecht (2002), Mars polar gully modification and possible formation from condensed volatiles, *Eos Trans. AGU*, *83*(47), Fall Meet. Suppl., Abstract P72C-10.

Byrne, S., and A. P. Ingersoll (2002), A sublimation model for Martian south polar ice features, *Science*, *299*, 1051–1053.

Carr, M. H. (1983), Stability of streams and lakes on Mars, *Icarus*, *56*, 476–495.

Christensen, P. R. (2003), Formation of recent Martian gullies through melting of extensive water-rich snow deposits, *Nature*, *422*, 45–48.

Costard, F., F. Forget, N. Mangold, and J. P. Peulvast (2002), Formation of recent Martian debris flows by melting of near-surface ground ice at high obliquity, *Science*, *295*, 110–113.

Davies, G. F., and R. E. Arvidson (1981), Martian thermal history, core segregation, and tectonics, *Icarus*, *45*, 339–346.

Dingman, S. L. (1984), *Fluvial Hydrology*, W. H. Freeman, New York.

Dingman, S. L. (1994), *Physical Hydrology*, Prentice-Hall, Upper Saddle River, N. J.

Doran, P. T., C. P. McKay, W. P. Adams, M. C. English, R. A. Wharton, and M. A. Meyer (1996), Climate forcing and thermal feedback of high arctic residual ice covers, *Limnol. Oceanogr.*, *41*, 839–848.

Edgett, K. S., M. C. Malin, R. M. E. Williams, and S. D. Davis (2003), Polar- and middle-latitude Martian gullies: A view from MGS MOC after 2 Mars years in the mapping orbit, *Lunar. Planet. Sci.*, *XXXIV*, abstract 1038.

Fanale, F. P. (1976), Martian volatiles: Their degassing history and geochemical fate, *Icarus*, *28*, 179–202.

Franck, S., and I. Orgzall (1987), A new view of Martian evolution, *Earth Moon Planets*, *37*, 287–313.

Gaidos, E. J. (2001), Cryovolcanism and the recent flow of liquid water on Mars, *Icarus*, *153*, 218–223.

Gilmore, M. S., and E. L. Phillips (2002), The role of aquicludes in the formation of Martian gullies, *Geology*, *30*, 1107–1110.

Golombek, M. P., and N. T. Bridges (2000), Erosion rates on Mars and implications for climate change: Constraints from the Pathfinder landing site, *J. Geophys. Res.*, *105*, 1841–1854.

Haberle, R. M., C. P. McKay, J. Schaeffer, N. A. Cabrol, E. A. Grin, A. P. Zent, and R. M. Quinn (2001), On the possibility of liquid water on present day Mars, *J. Geophys. Res.*, *106*, 23,317–23,326.

Hecht, M. H. (2002), Metastability of liquid water on Mars, *Icarus*, *156*, 373–386.

Heldmann, J. L., and M. T. Mellon (2004), Observations of Martian gullies and constraints on potential formation mechanisms, *Icarus*, *168*, 285–304.

Heldmann, J. L., W. H. Pollard, C. P. McKay, D. T. Andersen, and O. B. Toon (2005), Annual development cycle of an icing deposit and associated perennial spring activity on Axel Heiberg Island, Canadian High Arctic, *Arct. Antarct. Alpine Res.*, *37*(1), 126–134.

Jacobson, M. Z. (1999), *Fundamentals of Atmospheric Modeling*, Cambridge Univ. Press, New York.

Johnston, D. H., T. R. McGetchin, and M. N. Toksoz (1974), The thermal state and internal structure of Mars, *J. Geophys. Res.*, *79*, 3959–3971.

Kennard, E. H. (1938), *Kinetic Theory of Gases*, McGraw-Hill, New York.

Kirk, R. L., B. A. Archinal, E. M. Lee, M. E. Davies, T. R. Colvin, and T. C. Duxbury (2001), Geometric calibration of the Mars Orbiter Cameras and coalignment with the Mars Orbiter Laser Altimeter, *Lunar. Planet. Sci.*, *XXXII*, abstract 1863.

Knauth, L. P., and D. M. Burt (2002), Eutectic brines on Mars: Origin and possible relation to young seepage features, *Icarus*, *158*, 267–271.

Laskar, J., B. Levrard, and J. Mustard (2002), Orbital forcing of the Martian polar layered deposits, *Nature*, *419*, 375–377.

Malin, M. C., and K. S. Edgett (2000), Evidence for recent groundwater seepage and surface runoff on Mars, *Science*, *288*, 2330–2335.

Maxwell, J. B. (1982), *The Climate of the Canadian Arctic Islands and Adjacent Waters*, vol. 2, 589 pp., Atmos. Environ. Serv., Environ. Can., Gatineau, Quebec, Canada.

Mellon, M. T., and R. J. Phillips (2001), Recent gullies on Mars and the source of liquid water, *J. Geophys. Res.*, *106*, 23,165–23,179.

Mellon, M. T., B. M. Jakosky, H. H. Kieffer, and P. R. Christensen (2000), High-resolution thermal inertia mapping from the Mars Global Surveyor Thermal Emission Spectrometer, *Icarus*, *148*, 437–455.

Musselwhite, D. S., T. D. Swindle, and J. I. Lunine (2001), Liquid CO_2 and the formation of recent small gullies on Mars, *Geophys. Res. Lett.*, *28*, 1283–1285.

Pollard, W. H., C. Omelon, D. T. Andersen, and C. P. McKay (1999), Perennial spring occurrence in the Expedition Fiord area of western Axel Heiberg Island, Canadian High Arctic, *Can. J. Earth Sci.*, *36*, 105–120.

Schubert, G., and T. Spohn (1990), Thermal history of Mars and the sulfur content of its core, *J. Geophys. Res.*, *95*, 14,095–14,104.

Spohn, T. (1991), Mantle differentiation and thermal evolution of Mars, Mercury, and Venus, *Icarus*, *90*, 222–236.

Stevenson, D. J., T. Spohn, and G. Schubert (1983), Magnetism and thermal evolution of the terrestrial planets, *Icarus*, *54*, 466–489.

Toksoz, M. N., and A. T. Hsui (1978), Thermal history and evolution of Mars, *Icarus*, *34*, 537–547.

Toksoz, M. N., A. T. Hsui, and D. H. Johnston (1978), Thermal evolution of the terrestrial planets, *Moon Planets*, *18*, 281–320.

Treiman, A. H. (2003), Geologic settings of Martian gullies: Implications for their origins, *J. Geophys. Res.*, *108*(E4), 8031, doi:10.1029/2002JE001900.

Tschudin, K. (1946), Rate of vaporization of ice, *Helv. Phys. Acta*, *19*, 91–102.

Wallace, D., and C. Sagan (1979), Evaporation of ice in planetary atmospheres: Ice covered rivers on Mars, *Icarus*, *39*, 385–400.

D. T. Andersen and W. H. Pollard, Department of Geography, McGill University, Montreal, QC, Canada H3A 2K6.

J. L. Heldmann and C. P. McKay, Space Sciences Division, NASA Ames Research Center, Mail Stop 245-3, Moffett Field, CA 94035, USA. (jheldmann@mail.arc.nasa.gov)

M. T. Mellon, Laboratory for Atmospheric and Space Physics, University of Colorado, Boulder, CO 80309, USA.

J. Pitlick, Department of Geography, University of Colorado, Boulder, CO 80309, USA.

O. B. Toon, Program in Atmospheric and Oceanic Sciences, Laboratory for Atmospheric and Space Physics, University of Colorado, Boulder, CO 80309, USA.