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North-South Asymmetry in Martian Crater Slopes

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Abstract. The presence of an extensive ice-rich layer in the near subsurface of the Martian regolith can result in viscous creep responsible for softening craters at mid and high latitudes. The temperature of ground ice will vary spatially within a crater due the effect of slope on the effective angle of insolation. The temperature at a particular latitude will also vary temporally due to changes in Mars' obliquity. Results from numerical simulations of viscous creep indicate that these temperature variations cause the pole-facing slopes of craters to be systematically steeper than those of equator-facing slopes. Crater slopes should be most asymmetric between 25 and 40 degrees latitude, depending on the thickness of the creeping layer. This slope asymmetry predicted from theoretical simulations of regolith creep is not well developed in observed Martian crater topography. MOLA topography of craters 16 to 40 km in diameter was analyzed for north-south slope asymmetry within five latitude regions ranging from 60°S to 60°N. Based on the lack of any systematic slope asymmetry observed in the craters, we can place a conservative upper limit of $\sim$1.2 km on the thickness of the ice-rich creeping layer assuming a volumetric dust content of 70% and an exponentially increasing soil viscosity with depth. If the creeping layer contains relatively clean ice, then the thickness of ice-rich material is limited to $\sim$600 m or less based on our results. The observations also suggest that the thickness of this creeping layer is reduced by a factor of $\sim$2 towards the equator.

1. Introduction

The Martian subsurface is arguably the largest reservoir for H_2O on the planet. Based on thermophysical models, *Clifford* [1993] suggested that the Martian cryosphere could easily extend to 5 km depth globally. Given a surface porosity of 20%, a 5 km deep cryosphere amounts to a volume of water that could cover Mars in a 540 m deep global ocean. Constraining the thickness of the ice-rich regolith on Mars through geologic observations and remote sensing will allow for better estimates of the water budget.

Although direct evidence of water ice in the upper meter of the Martian subsurface has only recently been provided by neutron and gamma-ray spectrometer data and thermal emission observations [*Boynton et al.*, 2002; *Feldman et al.*, 2004; *Bandfield*, 2007], its presence had been strongly suspected since the Viking missions on both theoretical and observational grounds. Remote sensing observations [*Mustard et al.*, 2001] and numerical climate models [*Fanale et al.*, 1998; *Paige*, 1992; *Mellon et al.*, 1997] suggest that equatorial regions are generally depleted in near-surface ice, while an ice-rich mantle extends poleward of 30°.

To probe for ice at greater depths, we rely on remote sensing observations from radar reflectometry and on geological observations such as topographic roughness, crater softening, viscous flow features, and rampart ejecta. The only Mars radar results published to date are by *Plaut et al.* [2007] in which the thickness of the south polar layered deposits are measured at a thickness of 3 km. Future radar observations will likely constrain the thickness and distribution of ice-rich regolith at lower latitude. Global surface roughness measurements were made by *Kreslavsky and Head* [2003]. Their work suggests topographic relaxation has occurred at short wavelengths in mid-latitudes, possibly due to creep of ice-rich regolith. Ice-rich regolith creep may also be responsible for the observed north-south asymmetry in slopes at ~300 m length-scales [*Kreslavsky and Head*, 2000]. Relaxation of a deformable, presumably ice-rich, regolith layer of approximately 1 km in thickness may explain softened craters at mid to high latitude according to *Jankowski and Squyres* [1992]. The presence of lobate debris aprons and viscous flow features provide additional evidence for a large reservoir of ice-rich material in the near-surface of the Martian regolith [*Colaprete and Jakosky*, 1998; *Mangold et al.*, 2002; *Milliken et al.*, 2001]. Lobate ejecta deposits surrounding craters on Mars provide evidence for water ice in the near subsurface [*Strom et al.*, 1992]. *Kuzmin et al.* [1988, 1989] found a strong latitudinal dependence regarding the onset diameter of craters with rampart ejecta. They suggest that this latitudinal dependence reflects an enrichment of ice in the Martian regolith with increasing latitude.

Here we investigate crater relaxation as a probe of the thickness of the inferred subsurface ice-rich layer. Building on work by *Kreslavsky and Head* [2003] and *Jankowski and Squyres* [1992] we explore crater slope asymmetry as a means of determining the thickness of ice-rich regolith on Mars. First, we develop a theoretical model for crater slope asymmetry generation via viscous relaxation of crater topography. Next, we describe the method by which we analyzed real craters using Mars orbiter laser altimeter (MOLA) data [*Smith et al.*, 2001]. In the subsequent section we compare crater relaxation simulation results to the observations from various latitude regions on Mars in order to constrain the model parameters. Finally, we discuss the model and observational results regarding the thickness of ice-rich material in the Martian regolith and present our conclusions.

2. Theory

2.1. Downslope Creep of Ice-Rich Regolith

Downslope transport of ice-rich Martian regolith may be modeled as a diffusion creep process which smooths out topographic variations over time [*Colaprete and Jakosky*, 1998; *Perron et al.*, 2003; *Pathare et al.*, 2005]. Crater softening due to ice-driven creep may vary spatially within a single crater due to temperature variations induced by slope-related insolation variations. Because creep processes are more rapid at higher temperature, one would expect equator-facing crater slopes to creep more and become shallower than pole-facing slopes at low to moderate obliquities. The relaxation of topography (h) is governed by the thickness of the ice-rich layer, the viscosity at the surface (η_{sfc}) of the creeping layer, and the local slope. In order to minimize computation time, we make the simplifying assumption that the viscosity of the ice-rich Martian regolith increases exponentially with depth as follows:

$$\eta(z) = \eta_{sfc}\left(e^{z/\delta}\right) \tag{1}$$

where z is depth, increasing downward and the e-folding depth (δ) of viscosity will hereafter be referred to as the thickness of the creeping layer. Viscosity models presented by *Jankowski and Squyres* [1992] assumed either a constant viscosity with depth, or an exponentially decreasing viscsosity with depth. The latter of these models assumed the Martian geotherm had a dominant effect on the regolith rheology at depth. Here we assume that a decrease in the ice volume fraction with depth dominates the regolith rheology, resulting in an exponentially increasing regolith viscosity with depth. The viscosity at the surface of the creeping layer is given by

$$\eta_{sfc} = \eta_o \left[e^{\frac{Q(T-T_m)}{RTT_m} + b\phi} \right] \tag{2}$$

(modified from *Pathare et al.* [2005]) where Q is the activation energy, R is the ideal gas constant, T is the local temperature, T_m is 273 K, η_o is the viscosity of polycrystalline ice at 273 K, ϕ is the dust fraction, and b is a constant [*Durham et al.*, 1992]. Assuming the viscosity structure given above, a Newtonian rheology, and a constant ice-rich layer thickness, the following equation for the change in topography with time may be derived:

$$\frac{\partial h}{\partial t} = 2\delta^3 \rho g \frac{\partial}{\partial x} \left(\frac{1}{\eta_{sfc}} \frac{\partial h}{\partial x} \right) \tag{3}$$

where ρ is the density of the creeping layer and g is gravity. The derivation of equation 3 is given in Appendix A and the values of the parameters used in equations 2 and 3 are listed in Table 1. A viscosity of 10^{14} Pa s is appropriate for ice of 1.0 mm grainsize at 273 K (extrapolated from *Goldsby and Kohlstedt* [2001]). In our numerical simulations we consider ϕ values that range from 0 (dust-free ice) to the critical fraction at which the viscosity of a rock-ice mixture will be rock-dominated ($\phi = 0.72$) [*Mangold et al.*, 2002]. Table 1

2.2. Numerical Model

We modeled the relaxation of topography using a finite-difference discretization of equation 3 in which the only initial topographic variation is due to a single peak-to-peak sinusoidal curve. The peak-to-peak distance is the crater diameter (D), and the initial depth is set to $D/15$. This depth to diameter ratio was based on the maximum ratio observed in our analysis of crater topography and gives craters that are ~40% deeper than the depth to diameter relationship given by *Garvin and Frawley* [1998] for craters 30 km in diameter. The motivation for using a depth to diameter (d/D) ratio of 1/15 will be discussed in Section 5.2. Parabolic crater profiles were also used as an initial condition, but only sinusoidal results are presented in this paper for reasons that will be discussed in Section 5.1. The topographic profile, with the initial sinusoidal crater at the center, is divided into 1000 elements, each 100 m in length. The time step used in the simulation is 1000 yrs.

2.3. Influence of Obliquity and Slope on Temperature

The latitude, local slope, and obliquity all influence the local temperature. We determine the temperature of a horizontal surface on Mars at a given obliquity using theoretical insolation calculations by *Ward* [1974]. The annual average equatorial temperature on Mars is currently 220 K [*Mellon and Jakosky*, 1995]. This temperature corresponds to an insolation of 180 Wm^{-2} based on calculations by *Ward* [1974]. We can relate insolation values at other latitudes to temperature using the following relation:

$$T_{horiz} = \left(\frac{I}{180}\right)^{1/4} 220K \quad (4)$$

where T_{horiz} is the temperature of a horizontal surface and I is the local insolation in Wm^{-2} given by *Ward* [1974]. Figure 1 shows how temperature varies both as a function of latitude and obliquity in our simulations. In simulations lasting less than or equal to 10 Myrs, we allow obliquity to vary with time based on the recent obliquity history from *Laskar et al.* [2002]. For longer-term simulations, we use a statistical approach. The obliquity at a particular 1000 yr time step in the 100 Myr and 250 Myr simulations is randomly selected based on a gaussian distribution of obliquities given for the appropriate time into the Martian past (either 100 Myrs or 250 Myrs) determined by *Laskar et al.* [2004]. We assume the effect of slope is to change the local latitude by the angle of the slope. While

Figure 1

not exactly correct, this assumption is a good approximation of more complicated approaches (e.g. *Aharonson and Schorghofer* [2006]). For example, an equator-facing slope θ at a latitude ϕ is assigned a temperature of a latitude $(\phi - \theta)$. Before discussing the results from these simulations, we will first describe how we quantified slope asymmetry of craters in both the simulations and in the MOLA topography.

2.4. Slope Asymmetry

Based on the theory outlined above, we expect regolith creep to result in craters demonstrating a slope asymmetry with shallower equator-facing slopes than pole-facing slopes. We quantify the slope asymmetry in a similar fashion to *Kreslavsky and Head* [2003] by using the parameter

$$A = \frac{S_n - S_s}{S_{ave}} \tag{5}$$

where S_n and S_s are the maximum slopes on the north and south faces of the crater, respectively. In the simulations S_{ave} is the mean maximum slope for the two faces. A negative value for A indicates that the south face of the crater is steeper in slope than the north face. In the next section we describe the method used to measure crater slope asymmetry on Mars using gridded topography from MOLA.

3. Method of Crater Analysis

We used four criteria to select craters on Mars for slope asymmetry analysis. First, craters were chosen from five latitude bands, centered on: 60°N, 1.5°S, 20°S, 30°S, and 60°S. The locations of these regions are listed in Table 2. Second, we selected craters based on size, ranging from 16 to 40 km in diameter. Our analysis requires craters that are large enough to be resolved in gridded MOLA topography, and small enough so that no central peak is observed. Another criterion used in crater selection was freshness. We avoided overprinted craters, or craters with heavily modified rims. Finally, craters on steep regional slopes were not included in our survey to avoid potential bias in the calculation of A.

We measured crater slope asymmetry using 18 radial topographic profiles taken every 20 degrees of azimuth overlain on gridded MOLA topography at a pixel resolution of 460 m (Figure 2a). Each profile extends from the center of the crater to the crater rim, or just beyond the rim. The topography was interpolated at a resolution of 375 m for all profiles. Four profiles transect the north and south crater faces and 5 profiles transect the east and west faces. For each crater face we averaged the appropriate profiles together to get a stacked profile (Figure 2b). We then compared the maximum slopes from the north and south crater faces using the stacked profiles to determine slope asymmetry. The crater slope asymmetry is quantified by the parameter A, where $A = (S_n - S_s)/S_{ew}$ where S_{ew} is the average of the maximum slopes from the stacked east and west profiles. The quantification of slope asymmetry in crater topography, above, is slightly different than equation 5 because the topography is 3D and the simulations are done in 2D.

Figure 2

4. Results

We use a 100 Myr simulation time as an assumed minimum age of craters 16 to 40 km diameter on Mars. The crater chronology of Amazonis Planitia presented by *Hartmann and Neukum* [2001] suggest craters 16 to 40 km diameter give a surface age of ∼1 Gyr. Assuming a constant cratering rate, the average crater age would be 500 Myrs. Even though we are concentrating on the least degraded craters, we can assume most craters of this size are older than 100 Myrs. Running > 1 Gyr simulation times would be in-appropriate given the current uncertainties associated with long-term climate evolution on Mars. Results from 100 Myr simulations will give conservatively high estimates of the ice-rich creeping layer thickness if we are underestimating the true crater age. The majority of our simulations model 100 and 250 Myr of crater relaxation.

4.1. An Example Simulation

We calculate the theoretical change in topography of an initial crater profile over time using equations 1 through 4 with specified values for latitude, δ, and ϕ. The background surface temperature varies with time in the simulation due to the changing obliquity, as discussed in Section 2.3. An example of a modeled topographic profile is shown in Figure 3a for a 25 km diameter crater at 25° latitude with δ= 600 m and ϕ= 0.2 over a 10 Myr period. Notice the pronounced asymmetry in slope that has developed in the final profile due to the hastening of creep on the equator-facing slope (right hand side of crater). The value of $|A|$ in Figure 3a is 0.325. Figure 3b shows how the temperature along the initial and final profiles varies due to the effect of slope on the angle of incident sunlight. The shift in the regional temperature between the two profiles is due to a change in obliquity from 34.5° 10 Myr ago to 25° at present [*Laskar et al.*, 2002]. The model results shown in Figure 3 are simply examples; in general, simulations used an elapsed time of 100 Myr based on conservative estimates of likely crater ages (see above). Figure 3

4.2. 100 Myr Simulation Results

Figure 4 summarizes the result of 100 Myr simulations in which latitude and δ are varied. The simulation results show that at high latitudes surface temperatures are too cold and flow is so slow that no significant asymmetry develops. At mid latitudes the Figure 4 regional temperature is higher relative to the poles. Also, the temperature difference between the two slopes is significant due the differences in insolation between the north and south crater faces. The result is a local maximum in asymmetry at between 25° and 45° latitude, depending on the value for δ. Finally, near the equator, temperature is approximately the same for the north and south crater faces so no asymmetry can develop. As expected, the rate of flow and thus the development of asymmetry is a strong function of δ (equation 3). One important result from these simulations is the decrease in $|A|$ at low to mid-latitudes when the ice-rich layer thickness is increased from 600 m to 1000 m. This decrease in $|A|$ is a result of craters relaxing to the point that the opposing crater faces become more similar in slope with time. Thus, highly relaxed craters will have a lower value of $|A|$ than moderately relaxed craters. This effect will be discussed further in Section 4.5. Table 2.

4.3. Observational Results

The observed crater slope asymmetry in the five regions we analyzed is summarized in Table 2. In no case was a mean value of $|A|$ obtained that differed significantly from zero, indicating that the predicted systematic slope asymmetries (Figure 4) are not observed. Figure 5 compares the observed crater slope asymmetry to results from topographic relaxation simulations lasting 100 Myrs. The colored dots are measured slope asymmetries for individual craters. The black squares and vertical lines indicate the mean and $\pm$ 1 standard deviation in A, respectively, for craters in a given latitude region. As shown in Figure 5, there is no statistically significant slope asymmetry at any of the latitude regions we analyzed. This observation provides an upper limit on the extent to which creep has occurred. Since the creep rate depends mainly on δ and ϕ, the characteristics of any creeping ice-rich layer present can therefore be constrained. The black curves in Figure 5 show the expected asymmetry signal for a 30 km diameter crater evolving over a 100 Myr period for various creeping layer thicknesses with $\phi = 0.2$. The dashed line shows results from simulations with a higher dust fraction ($\phi = 0.7$) and a 1.2 km thick deformable layer. These results show that the mid-latitude slope asymmetry observations provide the strongest constraint on the creeping layer characteristics. The $\delta = 600$ m, $\phi = 0.2$ curve and the curve with $\delta = 1.2$ km and $\phi = 0.7$ fail to fit within one standard deviation of the observations. These combinations of ϕ and δ provide an upper bound on the creeping layer characteristics. As expected, there is a trade-off between δ and ϕ in generating crater slope asymmetry. As will be discussed in Section 4.5, longer timescale simulations will result in tighter constraints on δ and ϕ as one would expect.

Figure 5

4.4. Trade-offs Between δ, ϕ, and Simulation Time in Generating Slope Asymmetry

To further explore the trade-off between δ, ϕ, and creep duration we carried out a suite of runs lasting 10 and 100 Myr. These simulations used values for δ and ϕ that ranged from 200 m to 1000 m and from 0.2 to 0.8, respectively. Based on the observed slope asymmetries (Figure 5), runs that produced slope asymmetry values for mid-latitudes (40°) with $|A| < 0.2$ were deemed consistent with the observations. The tick-marked lines in Figure 6 show the domain of δ, ϕ, and creep duration that give a slope asymmetry value of 0.2 or less. As expected, a longer simulation duration results in tighter constraints on δ and ϕ. For a given value of $|A|$, higher values of ϕ permit larger thicknesses δ, and vice versa. For 100 Myr simulations with a dust fraction near the viscous transition to a solid particle rheology $\phi \approx 0.7$ [*Mangold et al.*, 2002] give a maximum layer thickness of

Figure 6

~ 1 km. A smaller dust fraction ($\phi = 0.2$) results in a correspondingly smaller maximum layer thickness (~ 0.35 km). For the 250 Myr simulations these maximum layer thicknesses are reduced to $\sim$0.8 km and $\sim$0.25 km, respectively.

4.5. 250 Myr Simulation Results

Our longest crater relaxation simulations lasted 250 Myrs. We chose reasonable values for δ and ϕ (1200 m and 0.5, respectively) based on the shorter simulations. Initial and final topography and slope of this simulation are shown in Figure 7a and b, respectively. The evolution of crater asymmetry (solid line) and depth to diameter ratio (dotted line) during the simulation is shown in Figure 7c. The reason for the decrease in A after 100 Myrs of simulation time is due to the shallow slopes present in a very relaxed crater. Once the crater has infilled beyond about one quarter of its original depth the slope asymmetry begins to decrease with time as shown in Figure 7c. This decrease in $|A|$ for highly softened craters is due to a reduction in the creep rate at low slopes (equation 3). After about one quarter of the crater has been infilled, the shallower of the two crater slopes experiences a slowed creep rate due to its low slope, allowing the creep process on the other crater face to catch up. As the crater topography continues to relax, the two slopes become more similar, resulting in a decrease in $|A|$. Because highly relaxed craters have a low $|A|$ value, 250 Myrs simulations with a large δ and small ϕ result in craters that satisfy the $|A| < 0.2$ agreement with the observations. However, as will be discussed later in this section, observed d/D ratios for the craters we analysed suggest highly relaxed craters are not found in the regions and size range that we analyzed.

Figure 7

The simulations were carried out using a time-dependent obliquity calculated using the statistical approach described in Section 2 (Figure 7d). In this figure we plotted obliquity every 100,000 yrs although it varied every time step (1000 yrs) in the simulation to make the plot readable. At high obliquity (>55°) the temperature of a horizontal surface gets warmer toward the poles [*Ward*, 1974] resulting in a reversal of the normal variation in creep rates. On average, however, the obliquity is about 34° over 100 to 250 Myr timescales [*Laskar et al.*, 2004]. Thus, during most of the numerical simulation, the temperature is decreasing toward the poles resulting in craters with a negative value for A in the northern hemisphere and a positive value in the southern hemisphere.

As the crater infills during our simulations, the diameter increases as the creep flow softens the crater rim. The combination of infilling and crater widening cause the crater's d/D ratio to significantly decrease over time depending on the creep rate. Comparing d/D ratios from the numerical model and the observations places an additional constrant on the creep rate. Figure 8a shows d/D ratio plotted against A for both simulation results and observations. Although the simulation is for a crater at 25°S latitude, observations from all the latitude regions we analyzed are plotted to illustrate the range in observed d/D ratio. The colored boxes give mean values for A and d/D ratio for a particular latitude region with ± 1 standard deviation given by the error bars. The simulation starts with a d/D ratio of 0.067 which decreases with time as the crater relaxes. The cross-hairs are separated by 25 Myr of simulation time; δ = 1200 m and ϕ = 0.5 in this 250 Myr simulation. The end-point of the simulation shown in Figure 8a has $|A| \approx 0.2$ and a d/D ratio much lower than the observed craters. The lack of low d/D ratio craters on Mars suggests few 16-40 km diameter craters have relaxed to the degree shown in Figure 7 or 8a. Figure 8

A d/D ratio versus A plot for a crater relaxation simulation at 60°S is compared to observations from 60°S and 60°N in Figure 8b. The negative of the mean A value for 60°N is plotted here in order to compare it with the southern latitude simulation. This simulation uses $\delta = 600m$ and $\phi = 0.5$. Figure 8c gives a similar plot for a simulation at 25°S with $\delta = 300$ m and $\phi = 0.5$. Observations from 30 and 20°S are shown for comparison. In both cases the end-point of the simulation (250 Myr) produces values of A and d/D ratio consistent with the observations.

5. Discussion

5.1. Summary of Results

The 100 Myr simulation results shown in Figure 6 suggest the ice-rich layer at 40° latitude could be either 300 m in thickness for clean ice or, alternatively, 1 km thick with a dust fraction of 0.7. Given the range of asymmetries in the observations, values for δ and ϕ are likely to be both spatially and temporally heterogeneous. At latitudes higher than 40°, crater relaxation is slowed by colder temperatures and a thicker creeping layer will result in a north-south slope asymmetry that still lies within the range of the observations.

One result from the 250 Myr simulations is that the age of the crater is important in determining the slope asymmetry that we observe today. Figure 7 shows that the maximum slope asymmetry occurs after about 100 Myrs of topographic relaxation. Subsequently, the slope asymmetry slowly decreases as more material infills the crater. The time at which the maximum asymmetry is reached depends on the size of the crater, the thickness of the creeping layer, the dust fraction, and latitude. The faster the creep of regolith, the sooner the maximum slope asymmetry will be reached. Based on the observed d/D ratios, the craters we analyzed on Mars have not relaxed so much that the slope asymmetry is decreasing with time (Figure 8a). Rather, the observed slope asymmetry is occuring in the early stage of crater softening. Figure 8b suggests, for a dust fraction of 0.5, that a creeping layer thickness of 600 m gives values of A and d/D ratio that are in close agreement with the observations at 60^o latitude. At lower latitude simulations with a creeping layer thickness of 300 m and $\phi = 0.5$ match the observed d/D ratios and A values (Figure 8c). A thickening creeping layer toward the poles is supported by cryosphere models put forth by *Clifford* [1993]; *Fanale et al.* [1998], and *Kuzmin et al.* [1989].

5.2. Model Assumptions

Our numerical model is less sophisticated than that used by *Pathare et al.* [2005] (although the basic physics is the same); the advantage of our simpler model is the ability to explore parameter space rapidly. We assume an exponentially increasing viscosity with depth attributed to the incorporation of more rocky material and a decrease in pore space (equation 1). We also assume a linear relationship between shear stress and shear strain rate. Although ice can deform in a non-Newtonian manner, debris apron modeling by *Mangold and Allemand* [2001] suggests that assuming a Newtonian rheology may be appropriate. Ice is assumed to be present at/within the surface during the simulations, extending to a constant depth scaled according to the creeping layer thickness. As described in *Mellon and Jakosky* [1995], surface ice is currently unstable at latitudes less than $\sim$60^o, resulting in a dessicated layer extending a few meters into the Martian regolith. However, these authors also show that surface ice is stable at all locations on Mars at obliquities $> 32^o$. The thickness of the dessicated layer is insignificant compared to the creeping layer thicknesses ($\sim$ 1 km) used in our simulations, and will have a negligible effect on its rheology. The presence of ice deep within the Martian subsurface has been suggested by *Clifford* [1993], extending to perhaps 10 km. In this paper we have not accounted for where ice is, or was, stable in the Martian subsurface, rather we have simply modeled how craters would respond to ice-rich regolith creep assuming that it is present. One should keep in mind, however, that the instability of ice at the surface does not necessarily correlate with a lack of ice at depth.

Many of the assumptions used our numerical model were made to ensure that our estimate of the ice-rich layer thickness was an upper limit. For instance, we use a sinusoidal initial topographic profile instead of a parabolic profile because the slopes of the parabolic profile are more steep and result in a larger slope asymmetry as the parabolic profile begins to relax. In general, a larger creeping layer thickness generates more asymmetric crater slopes as long as the crater doesn't become significantly infilled (Figures 4 and 6). Using a sinusoidal initial profile forces the creeping layer to be thicker than it would be for a parabolic profile in order to produce the same slope asymmetry. Another conservative assumption used in our model is the viscosity at the surface of the ice-rich creeping layer (η_o) and the vertical viscosity structure given by equation (2). The 10^{14} Pas surface viscosity is for rather coarse-grained ice (1.0 mm) giving a higher creeping layer viscosity than would be present for finer grained ice. Assuming a relatively high viscosity slows the rate of crater relaxation - making our estimate for the creeping layer thickness an upper limit. We also limit the rate at which the craters relax by assuming an exponentially increasing viscosity with depth. In order to reproduce a softened crater, the creeping layer thickness must be greater using our viscosity structure than it would be for a viscosity structure with smaller depth variations or a lower surface value (smaller ice grain size). Finally, assuming a relatively large d/D ratio (compared to *Garvin and Frawley* [1998]) ensures that our creeping layer thickness is an upper limit. Although a larger d/D ratio results in steeper slopes, the creeping layer thickness must still be greater in order to produce a softened crater with a d/D ratio that lies within the range of observations (Figure 8).

Although we have accounted for obliquity variations in our simulations, we did not address the effect of a time dependent eccentricity, nor variations in the longitude of periapse. Both of these effects will have a smaller effect on the temperature distribution than obliquity variations [*Laskar et al.*, 2002].

5.3. Additional Remarks

There remain many questions regarding the mechanism by which craters are softened on Mars. For instance: why are no systematic crater slope asymmetries observed in any of the regions we studied on Mars, especially in the high latitude regions?

One possibility is that the ice-rich layer thickness varies spatially within a crater. At low to moderate obliquities the pole-facing crater slope is colder than the equator-facing slope. However, creep hastening on the warm slope might be subdued if the ice-rich layer is thin. If coupled with a thick creeping layer on the cold slope, the creep rates for the north and south slopes could be approximately equal - resulting in no slope asymmetry even though topographic relaxation proceeds. The preferential surface deposition of ice-rich mantling material on pole-facing slopes [*Aharonson and Schorghofer*, 2006] might also affect the topography and reduce the effective north-south topographic asymmetry. However, the thickness of mantling material required to change the slope on a 16 km diameter crater by 5^{o} is about 700 m. This thickness seems unlikely given the 10 m mantling thickness estimates provided by *Mustard et al.* [2001]. One mechanism that does not appear able to explain the observations is temperature variations induced by changes in obliquity. Although obliquity variations may affect topography at ~ 300 m lengthscales [*Kreslavsky and Head*, 2003], our simulations (e.g. Figure 7) suggest that obliquity variations are not sufficient to override the formation of a north-south slope asymmetry in craters 16 to 40 km in diameter.

6. Conclusions

Topographic observations of craters on Mars indicate that there is no statistically significant dependence of crater slope asymmetry on latitude (Section 4.3, Table 2). Generally, slope asymmetry ranges between $A = -0.2$ and $A = 0.2$ in all of the five regions we analyzed. Numerical models of crater relaxation using reasonable creeping layer thicknesses and dust fractions suggest that the formation of significant slope asymmetry occurs over $\sim$100 Myr timescales. Comparison between topographic observations of craters on Mars and our numerical models suggest that a clean ice-rich layer on Mars does not extend deeper than 600 m. If the creeping layer responsible for crater softening on Mars contains 50 - 70% dust per volume, then the creeping layer thickness could extend to 1200 m. At low latitude the creeping layer is likely thinner, perhaps 300 m or less, based on d/D ratio comparisons between the simulations and the observations (Figure 8). These thickness estimates are conservative and would be lower if, for example, longer simulation times were used. Assuming a 600 m thick global near-surface, ice-rich layer incorporating a dust fraction of 0.2 accounts for $\sim 6 \times 10^7$ km^3 of water. This upper-limit volume estimate is a factor of 3 smaller than the Noachian ocean volume proposed by *Clifford and Parker* [2001], but is in agreement with the end-state of the hydrosphere evolution model that these authors propose. The extent of the northern ocean proposed by *Baker et al.* [2000]

(reaching Olympus Mons) exceeds our upper limit water volume estimate by a factor of 5. More modest-sized oceans [*Parker et al.*, 1989, 1993; *Carr and III*, 2003], extending to the outer Vastitas Borealis formation, are less than half of our estimated volume.

It is unclear why the mean crater slope asymmetry in a particular region is so close to zero. Most puzzling is the number of craters with values of A that are opposite in sign to what one would expect theoretically (Figure 5). A range of crater asymmetry values (of a given sign) within a particular region on Mars is expected because crater slope asymmetry evolves over time (Figure 7c). We expect there to be some scatter in A due to the range of crater ages present within a particular region, but not a variation in sign as a result of creep-related processes. One possibility is simply that regolith creep is sufficiently slow (e.g. the creeping layer is thin) that this process is overwhelmed by other geological processes such as gully formation or tectonic deformation.

The strong dependence of crater softening on wavelength [*Jankowski and Squyres*, 1992; *Pathare et al.*, 2005] is another possible explanation of why no systematic crater asymmetries are observed. Because short wavelength features will relax more quickly than longer wavelengths, it is possible that smaller craters than those we have examined will show systematic crater slope asymmetry. Alternatively, an increase in the creeping layer thickness on cold slopes coupled with a reduction of ice-rich material on warm slopes would significantly reduce the expected crater slope asymmetry.

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7. Appendix A. Derivation of the Momentum Equation

Here we derive equation 3, which describes the relaxation of topography resulting from the flow of ice-rich regolith downslope. We assume an exponentially increasing viscosity with depth (equation 1), a Newtonian rheology, a constant e-folding viscosity depth δ, and conservation of mass. The shear stress on a deforming viscous medium is given by:

$$\frac{\tau}{\eta} = \frac{\partial v}{\partial z} \tag{6}$$

where τ is shear stress, η is the Newtonian viscosity and v is down-dip velocity. We choose a coordinate system in which v is oriented down-dip and z is perpendicular to the sloping surface, increasing downward. Topographic gradients result in shear stresses (given in differential form):

$$-\frac{\partial \tau}{\partial z} = \rho g \frac{\partial h}{\partial x} \tag{7}$$

where g is gravity, ρ is density of the ice-rock mixture, and h is elevation. Integrating equation (7) and substituting into (6) results in the following equation:

$$\frac{\partial v}{\partial z} = (\eta_{sfc})^{-1} e^{-z/\delta} \left(\rho g \frac{\partial h}{\partial x} z + c \right) \tag{8}$$

where c is a constant of integration and η_{sfc} is given by equation 2. Using $\partial v/\partial z = 0$ at $z = 0$ as a boundary condition c becomes 0. Integrating once more and applying the boundary condition that $v = 0$ at $z = \infty$ gives the expression for the vertical velocity field of downslope ice-rich regolith creep

$$v(z) = \frac{\rho g}{\eta_{sfc}} \frac{\partial h}{\partial x} \left[\delta e^{-z/\delta} (z + \delta) \right]. \tag{9}$$

To translate this velocity field into a change in topography, we use the following conservation of mass relationship:

$$\frac{\partial h}{\partial t} = -\frac{\partial}{\partial x}\int_0^\infty v \partial z. \tag{10}$$

Finally, we combine equations (9) and (10) to obtain the momentum equation given in (3).

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Figure 1. Local temperature of a horizontal surface on Mars as a function of latitude for the indicated obliquity state. Calculated using the method described in *Ward* [1974] and Section 2.3.

Figure 2. a) MOLA gridded topography of a 22 km diameter crater at 58.0ºS, 156.5ºE. Topographic contours are made every 100 m in elevation. Dotted lines indicate locations of radial profiles used to measure slope. b) Averaged topographic profiles for the northern crater face (dashed line) and the southern crater face (solid line). This crater has an asymmetry of $A = 0.35$ (north face is steeper).

Figure 3. a) Simulation of 10 Myrs of topographic relaxation of a 25 km diameter crater at 25º latitude assuming a deformable layer 600 m in thickness with a 20% dust fraction. The final crater has a slope asymmetry of $|A| = 0.3250$. b) The temperature along the initial and final profiles varies due to the effect of slope on the angle of incident sunlight. The shift in the regional temperature between the two profiles is due to an obliquity of 34.5º 10 Myr ago compared to 25º present obliquity [*Laskar et al.*, 2002].

Figure 4. Numerical results of the crater asymmetry parameter ($|A|$) as a function of latitude and δ. These results assume a 30 km diameter crater, a 20% dust fraction, a 100 Myr relaxation period, and the indicated deformable layer thickness δ.

Figure 5. Observed crater asymmetry measured at various latitudes (colored dots) compared to theoretical results (black curves) for varying creeping layer thicknesses (δ) with ϕ=0.2 over a 100 Myr period. The black squares and vertical lines indicate the mean asymmetry values and $\pm$ 1 standard deviation, respectively, for a particular latitude region. The dashed line shows results from simulations with a high dust fraction ($\phi = 0.7$) and a 1.2 km thick deformable layer.

Figure 6. Asymmetry contours of $|A| = 0.2$ given values for δ, ϕ, and simulation duration. A crater diameter of 30 km, and a latitude of 40° were used in these simulations. The domain of $|A| \leq 0.2$ for each simulation duration is indicated by the tick marks.

Figure 7. a) Simulation of 250 Myrs of topographic relaxation of a 25 km diameter crater at 40° latitude assuming a deformable layer 1200 m in thickness with a 50% dust fraction. The final crater has a slope asymmetry of $|A| = 0.3973$. b) The slope along the initial and final topographic profiles. c) The change in slope asymmetry (A) over time. d) Obliquity variation over time using the statistical approach described in Section 2.3. We plotted obliquity every 100,000 yrs although it varied in the simulation every time step (1000 yrs).

Figure 8. a) Simulation results and observations of depth to diameter (d/D) ratio plotted against A. This simulation lasts 250 Myr and is for a crater at 25°S latitude. The cross-hairs are separated by 25 Myr of simulation time; δ=1200 m and ϕ=0.5 in this simulation. The colored boxes give observed mean values for A and d/D ratio for a particular latitude region with $\pm$ one standard deviation given by the error bars (Table 2). b) d/D ratio versus A plot for a crater relaxation simulation at 60°S using a δ of 600 m and ϕ of 0.5. Observations from 60°S and the negative of the mean A value for 60°N are plotted here in order to compare it with the southern latitude simulations. c) Similar plot for a simulation at 25°S with δ=400 m and $\phi = 0.5$. Observations from 30 and 20°S are shown for comparison.

Table 1. Definitions and measured or theoretical values (or range of values) for parameters used in the numerical simulations

Description	Symbol	Value(s)	Reference
activation energy	Q	50 kJ mol^{-1}	[*Goldsby and Kohlstedt*, 2001]
reference viscosity	η_o	10^{14} $Pa\ s$	[*Goldsby and Kohlstedt*, 2001]
dust fraction	ϕ	0.2−0.72	(upper limit) [*Mangold et al.*, 2002]
dust frac. coeff.	b	8	[*Pathare et al.*, 2005]
density of creep layer	ρ	1.07−2.01 g cm^{-3}	based on ϕ, above
gravity	g	3.7 m s^{-2}	

Table 2. Crater statistics for each region based on topographic observations [a]

Latitude	Longitude	# of Craters	A	Mean Diam (km)	Mean Depth (km)
58−63° N	0−360° E	27	0.02±0.19	26.1±6.4	1.01±0.41
0.25−2.75° S	30−70° E	11	−0.17±0.39	28.0±8.7	0.67±0.29
18−22° S	330−360° E	13	−0.155±0.28	29.8±8.1	1.15±0.34
28−35° S	130−150° E	23	−0.134±0.29	29.6±8.9	1.16±0.56
58−63° S	130−170° E	23	−0.07±0.25	29.9±7.1	0.91±0.36

[a] Errors are ± 1 standard deviation.