IMPACT-INDUCED MANTLE DYNAMICS ON MARS. James H. Roberts, *Johns Hopkins University Applied Physics Lab, Laurel, MD 20723 (James.Roberts@jhuapl.edu)*, Jafar Arkani-Hamed, *University of Toronto, Department of Physics, Toronto, ON, Canada M5S 1A7*.

Introduction: At least 20 exposed and buried impact basins with diameters D_b exceeding 1000 km have been identified on Mars [1]. The five largest of these have $D_b > 2500$ km, approaching the radius of the planet. The crater retention ages (CRAs) of these basins show a strong clustering, implying a peak in crater production between 4.2 and 4.1 Ga [2], that may be related to the Late Heavy Bombardment [3].

Mars currently has no global magnetic field, however widespread crustal magnetization [4] provides strong evidence that such a field existed in the past. The crustal magnetization for the giant basins is very strongly correlated with their ages. The older basins are all strongly magnetized; the younger basins are all de-magnetized. Furthermore, there is a monotonic decrease over time in magnetic field strength measured over the five largest basins [5]. The absence of magnetization in the younger basins suggests that a dynamo operated during the early to mid-Noachian [6], but stopped once the heat flow became unfavorable for core convection. The coincidental timing of the end of the sequence of impacts and the disappearence of the global magnetic field has led to investigations of a causal connection between the two events [8-9].

The rate of core cooling (and thus dynamo activity) is limited by the cooling rate of the overlying mantle. Thus, the extent to which impacts can affect the core depends upon the pre-existing thermal state of the mantle. Futhermore, the ability of the CMB heat flow (and potentially the dynamo) to recover from an impact depend on the time for the impact heating to dissipate relative to the frequency of impactors. Here, we examine the effects of the initial thermal structure of the core and mantle, the location of an impact with respect to the pre-existing convective structure, and on the mantle dynamics and CMB heat flux. We also examine the cumulative effects of multiple impacts.

Convection Model: We model thermal convection in the Martian mantle using the 3D spherical finite-element convection code CitcomS [10] with the extended Boussinesq approximation [11]. The viscosity of the mantle is temperature- and pressure-dependent, following an Arrhenius-style law with activation parameters chosen to be consistent with the visocity of the Earth's upper mantle [12]. We apply temperature and free-slip boundary conditions at the surface and core-mantle boundary (CMB). The mantle is heated from within by time-dependent radioactive decay, and by secular cooling of the core. We ran several models with different initial mantle temperature profiles and CMB temperatures.

Impact Heating: Impacts can introduce a substantial amount of heat into the interior of the planet. We use scaling relations to obtain the transient basin diameters from the observed final basin sizes, D_b [13], and to obtain the impactor size from the transient basins [14], assuming an impactor velocity of 10 km/s. A significant fraction of the impactor's kinetic energy will be converted to thermal energy, raising the temperature of the surrounding mantle. The mantle is heated by a shock wave emanatimg from the impact location. Heating is uniform within an isobaric core, which scales with the impactor size and decays rapidly outside this region [15]. We parameterize the impact heating as a temperature perturbation in the mantle, which is a function of the shock pressure [16].

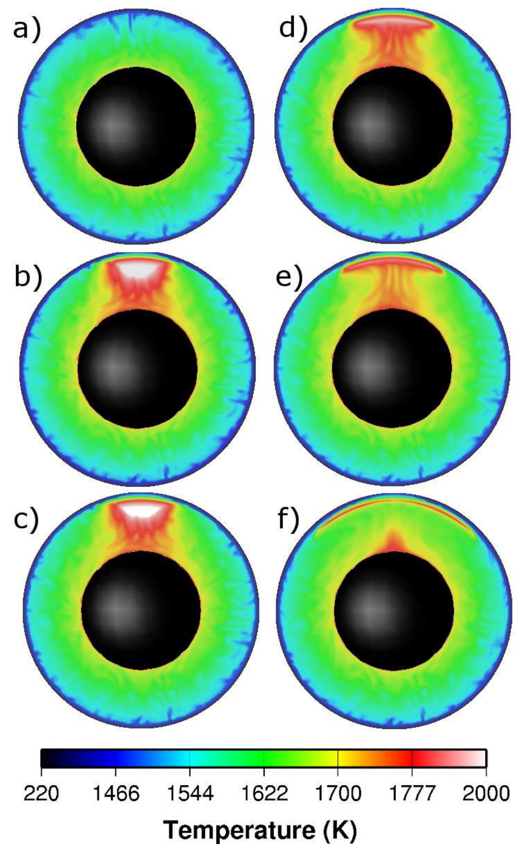

Figure 1: Temperature structure on cross-section of the mantle though the impact site immediately before (a) and after (b) the impact, after 3 ky (c), 100 ky (d), 140 My (e), and 1.8 My (f)

Results: At the time of formation [2] of the Utopia basin, we impose the appropriate impact heating due on our standard background case. Snapshots of the temperature structure are shown in Figure 1 for a cross-section through the impact site. The mantle is initially convecting prior to the impact with multiple upwellings, and small-scale downwellings (Figure 1a). The impact heats a region of the mantle. Heating is strongest within a truncated sphere-shaped isobaric core and decays with distance (Figure 1b). This heated region rises and spreads out, forming a "thermal blanket" that insulates the mantle (Figure 1c,d). An upwelling develops at the CMB, rises, and punctures the thermal blanket (Figure 1e,f). The

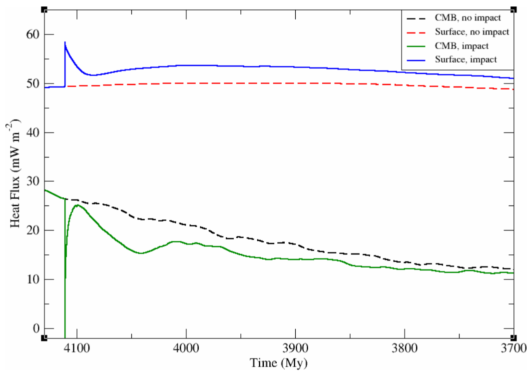

Figure 2: Evolution of the heat f ux at the surface and CMB for the model shown in Figure 1 with (solid) and without (dashed) a Utopia-forming impact.

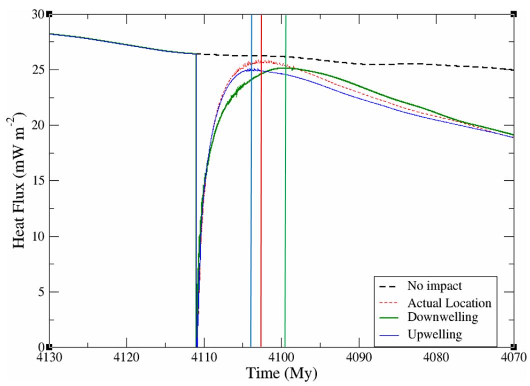

Figure 3: CMB heat f ow evolution in response to Utopia-forming impacts at different locations with respect to the pre-existing convective structure. Vertical lines mark the times of initial heat f ow recovery for each model.

blanket ultimately dissolves, leaving a degree-1 convective pattern with the upwelling beneath the impact site. Compared with the background model, the impact heating causes a rapid increase in surface heat f ux and corresponding decrease in CMB heat f ux as core cooling is restricted. The resulting buoyancy and spreading of the heated region results in fairly rapid dissipation of the heat f ux anomaly. Within about 10 My, the heat f ux at both boundaries is nearly restored to the pre-impact level. However, total recovery does not occur. The change in convective planform results in a long-term elevation (decrease) in surface (CMB) heat f ux of a few percent. This heat f ux evolution is shown in Figure 2

We have investigated the effects of pre-existing convective structure on the response to impact heating. Here we placed a Utopia-forming impactor over a convective upwelling, over a downwelling, and over a quiescent region. Figure 3 shows the evolution of the CMB heat f ux for each of these cases. The behavior is similar in all cases except for the timing. The pre-existing convective motion can either aid or hinder the recovery. Impact heating over an upwelling will be removed about 40% faster than over a downwelling. In all cases, however, there is some long-term alteration to the heat f ow beyond the initial recovery period.

Finally, we investigated the cumulative effects of multiple impacts on the mantle dynamics. We took from [2] the times and locations of the twenty largest impacts and imposed their impact heating. Only the very largest impacts ($D_b > 2500$ km) can signif cantly heat the lower mantle. Therefore, we ran a similar model with only the f ve largest impacts. The evolution of heat f ux for these models is shown in Figure 4. We f nd that the top f ve impacts account for the majority($\sim$ 80%) of the heat f ow anomalies. The mean time between the largest impacts is around 25 My, which is longer than the recovery time for a single impact. However, the heat f ow anomalies grow with each successive impact, suggesting a cumulative effect of the long-term change in heat f ow from each impact.

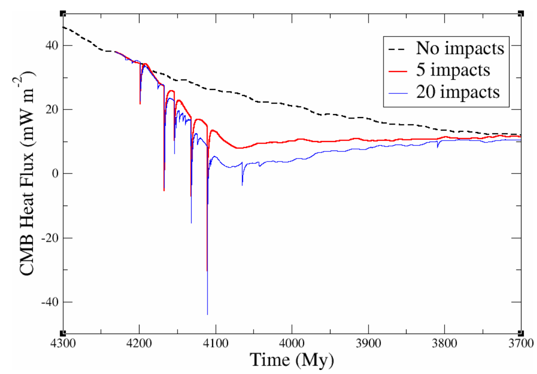

Figure 4: Surface (top) and CMB (bottom) heat f ow vs. time for a sequence of the 20 (blue) and 5 (red) largest Noachian impacts.

Conclusions: We f nd that a single basin-forming impact can alter the entire f ow f eld of the mantle. The impact results in the formation of a long-wavelength upwelling beneath the impact site which sets up a persistent degree-1 convective pattern in the mantle. The interval between the largest impacts is longer than the initial recovery time for a single impact. However, the change in convective pattern due to each impact sets up a long term change in the global heat f ow. These long-term changes are cumulative, and multiple impacts have a greater effect than a single impact. We note also that only the very largest impacts can heat the lower mantle. Thus the effects of the top 20 basins can be approximated by the top 5.

References: [1] Frey, H. V. (2008) *GRL, 35*, L13203. [2] Hartmann, W. K. and Neukum, G. (2001), *Space Sci. Rev., 96* 165-194. [3] Cohen, B. A. et al. (2000), *Science, 290* 1754-1755. [4] Acuna, M. H. et al. (2001) *JGR, 106*, 23,403-23,417. [5] Lillis, R. J. et al. (2008) *GRL, 35*, L14203. [6] Arkani-Hamed and Boutin (2004) *JGR, 109*, E03011. [7] Nimmo F. and Stevenson D. J. (2000) *JGR, 105*, 11,969-11,979. [8] Roberts, J. H. et al. (2009), *JGR, 114*, E04009. [9] Arkani-Hamed, J. and Olson, P. (2010) , *JGR, 115* E07012. [10] Zhong, S. et al. (2000), *JGR, 105*, 11063-11082. [11] Christensen, U. and Yuen, D. A. (1985) *JGR, 90*, 10,291-10,300. [12] Mitrovica, J. X. and Forte, A. M. (2004) *EPSL, 225*, 177-189. [13] Holsapple, K. A. (1993) *AREPS, 21* 333-373. [14] Melosh, H. J. (1989) *Impact cratering*, Oxford Univ Press, 253 pp. [15] Pierazzo, E. et al. (1997), *Icarus, 127* 208-223. [16] Watters, W. et al. (2009) *JGR, 114*, E02001.