Stagnant Lid Heterogeneity on Mars

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Introduction: The mantle of Mars is in the stagnant lid convection regime where most of the cold upper boundary layer is immobile and conductive. It is shown here how crust thickness can influence the thickness and heat flow of the lid.

Models of the crustal structure of Mars, derived using topography data from the Mars Orbiter Laser Altimeter (MOLA) and gravity from the Radio Science investivgations of the Mars Global Surveyor (MGS) spacecraft [1][2][3], provide estimates of the global crustal thickness. The crust thickness is characterized by a roughly hemispheric (degree 1) structure that represents the geophysical manifestation of the crustal dichotomy (Fig. 1). Assuming a constant crust density, ρ_c = 2900 kg m^{-3}, mantle density, ρ_m = 3500 kg m^{-3}, and an average global crust thickness, t_{cr} = 45 km, Neumann et al. [2] find the dichotomy corresponds to average crust thicknesses of 32 km and 58 km in the northern lowlands and southern highlands repectively. This 26 km difference in average crust thickness between the two terrains will result in differences in bulk lithospheric properties. For example, Grott and Breuer [4] demonstrate how effective elastic thickness, T_e, can be influenced by crustal structure.

Heat flow on Mars can be estimated from the modeled effective elastic thicknesses which quantify the characteristic wavelength of elastic flexure of the lithosphere from observed gravity and topography. The polar ice caps provide the only estimate of the present-day heat flow on Mars as these formed in the recent geologic past (e.g. [5][6]). The maximum $\sim$100 m deflection of the lithosphere in response to the ice load of the north polar cap implies that $T_e \geq 300\ km$ presently [7].

Model: Taking $T_e = 300$ km provides a lower limit on the present-day stagnant lid thickness beneath the north polar cap. For lithosphere with minimal deflection, T_e corresponds to the base of the mechanical lithosphere, the depth at which the lithosphere can no longer store stress on geologic time-scales and is effectively dissipated by ductile deformation. In this case, T_e can be ascribed to the depth of an isotherm. At 300 km, $T = 1233$ K assuming a wet olivine rheology [8] and a strain rate of 10^{-14} s^{-1} appropriate for a the loading time scale of the ice cap. Temperature profiles in the stagnant lid are then constructed using an average cap surface temperature, $T_s = 155$ K, and thickness of 3 km, crust thickness $t_{cr} = 32$ km, a crustal heat production rate $H_c = 4.91 \times 10^{-5}$ μW kg^{-1} from the decay of K, Th, and U [9], mantle heat production $0.1 H_c$, and a temperature dependent conductivity in the mantle [10]. The surface heat flow that yields $T = 1233$ K at 300 km depth is found to be $F_s = 16.6$ mW m^{-2}. Taking a representative mantle interior temperature of 1600 K, the base of the stagnant lid above the upper thermal boundary layer of the convecting mantle occurs at a depth of $D = 349$ km and the heat flow from the conductive interior is $F_m = 6.5$ mW m^{-2}.

Stagnant Lid Parameterization: Convection in the mantle interior is nearly isoviscous and driven by the small temperature difference across the rheological sublayer at the base of the stagnant lid which depends on the rheological law of the mantle and the mantle's internal temperature (e.g. [11][12]). Convective instabilities are determined by the local conditions in this layer, and therefore heat flow is independent of surface temperature or the influence of the overlying crust on the thermal gradient in the stagnant lid. Differences between heat flow from the convecting mantle and heat flow through the conductive lid however will result in changes in the lid thickness. For example, as heat is lost from the interior over time, the depth at which the interior experiences the viscosity contrast forming the rheologic sublayer, occurs at greater depth, and therefore the stagnant lid thickens. A change in crust thickness alters the thermal gradient in the lid due to a different distribution of heat producing elements with a thicker crust resulting in higher Moho temperatures and a shallower thermal gradient in the underlying mantle portion of the lid. To compensate, the lid thickness will decrease to to balance the heat flow from the underlying convecting mantle.

Results: Given F_m, the stagnant lid thickness can be determined for other regions of the planet with different crust thicknesses for the assumed mantle interior temperature. To determine the magnitude of the of the stagnant lid thickness variation across the crustal dichotomy boundary, the lid thickness for the northern lowlands is determined, now without an ice cap and using the average global surface temperature ($T_s = 220$ km), and the southern highlands with an average crust $\sim$26 km thicker. Initial results yield $F_s = 16.4$ mW m^{-2} and $D = 338$ km for the northern lowlands and $F_s = 18.9$ mW m^{-2} and $D = 295$ km for the southern highlands.

The interior of Hellas Basin represents some of

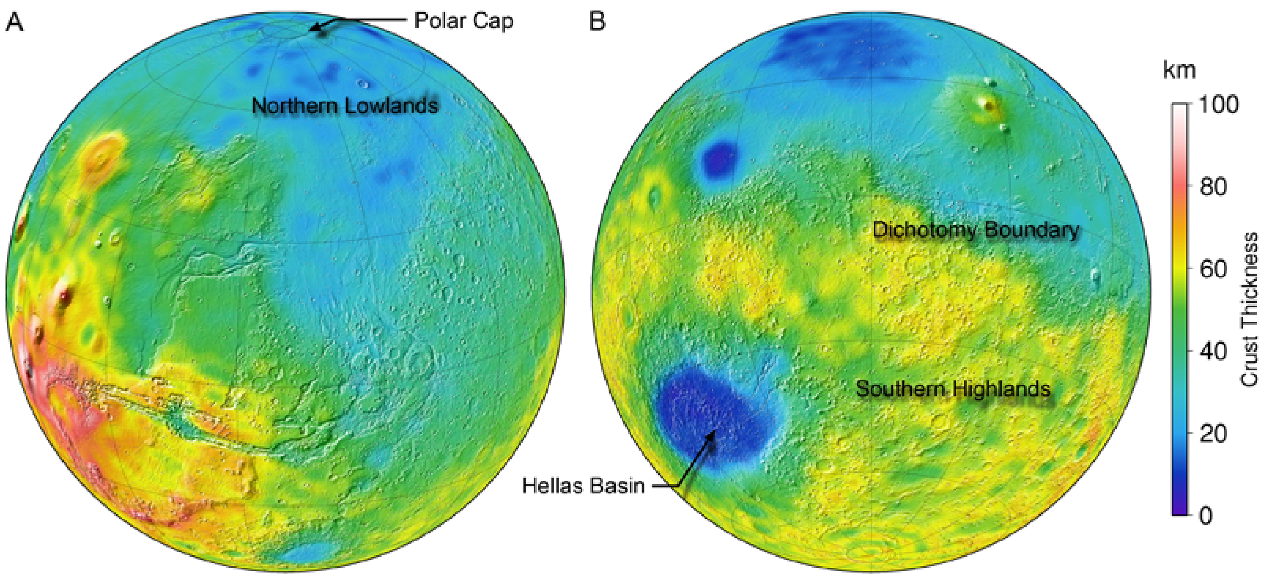

Figure 1: Crust thickness of the (A) western hemisphere and (B) eastern hemisphere of Mars from[2].

the thinnest crust on Mars, and along with Argyre basin, is the only crust thinner than 30 km in the southern highlands. Taking $t_{cr} = 7$ km as a typical value for the basin interior, gives $F_s = 14.0$ mW m^{-2} and $D = 382$ km.

Discussion: Using the crust thickness model of Neumann et al. [2] and adopting typical values characterizing stagnant lid parameterization, a present-day stagnant lid thickness difference of 43 km between the northern lowlands and the southern highlands is implied. The thicker crust in the highlands would result in a slower rate of stagnant lid growth as the interior cooled. Such a degree 1 feature in the rigid upper boundary of the deeper convecting interior has implications for mantle convection including possible edge driven convection and lithospheric delamination that could influence the location of volcanism. Further, Hellas represents a 87 km bulge in the base of the lid in the southern hemisphere which could interact with and influence the pattern of mantle convection by nucleating small-scale convective instabilities.

References

[1] M. T. Zuber, *et al.* Internal structure and early thermal evolution of Mars Global Surveyor topography and gravity. *Science*, 287:1788–1793, 2000.

[2] G. A. Neumann, *et al.* Internal structure and early thermal evolution of Mars Global Surveyor topography and gravity. *J. Geophys. Res.*, 109, 2004. doi: 10.1029/2004JE002262.

[3] M. A. Wieczorek and M. T. Zuber. Thickness of the Martian crust: Improved constraints from geiod-to-topography ratios. *J. Geophys. Res.*, 109, 2004. doi: 10.1029/2003JE002153.

[4] M. Grott and D. Breuer. On spatial variability of the Martian elastic lithosphere thickness: Evidence for mantle plumes? *J. Geophys. Res.*, 115, 2010. doi:10.1029/2009JE003456.

[5] K. E. Fishbaugh and J. W. Head. Comparison f the north and south polar caps of Mars: New observations from MOLA data and discussion of some outstanding questions. *Icarus*, 154:145–161, 2001.

[6] B. Levard, *et al.* Recent formation and evolution of narthern Martian polar layered deposits as inferred from a Global Climate Model. *J. Geophys. Res.*, 112, 2007. doi:10.1029/2006JE002772.

[7] R. J. Phillips, *et al.* Mars north polar deposits: Stratigraphy, age, and geodynamical response. *Science*, 320:1182–1185, 2008. doi:10.1029/science.1157546.

[8] S. Karato and P. Wu. Rheology of the upper mantle: A synthesis. *Science*, 260:771–778, 1993.

[9] B. C. Hahn, *et al.* Martian surface heat production and crustal heat flow from Mars Odyssey Gamma-Ray spectrometry. *Geophys. Res. Lett.*, 38, 2011. doi: 10.1029/2011GL047435.

[10] D. McKenzie, *et al.* Thermal structure of oceanic and continental lithosphere. *Earth. Plant. Sci. Lett.*, 233:337–349, 2005.

[11] V. S. Solomatov. Scaling of temperature- and stress-dependent viscosity convection. *Phys. Fluids*, 7:266–274, 1995.

[12] O. Grasset and E. M. Parmentier. Thermal convection in a volumetrically heated, infinite prandtl number fluid with strongly temperature-dependent viscosity: Implications for planetary thermal evolution. *J. Geophys. Res.*, 103:18171–18181, 1998.