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Implications of large elastic thicknesses for the composition and current thermal state of Mars

M. Grott *, D. Breuer

Institute of Planetary Research, German Aerospace Center (DLR), Rutherfordstraße 2, 12489 Berlin, Germany

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ABSTRACT

The martian elastic lithosphere thickness T_e has recently been constrained by modeling the geodynamical response to loading at the martian polar caps and T_e was found to exceed 300 km at the north pole today. Geological evidence suggests that Mars has been volcanically active in the recent past and we have reinvestigated the martian thermal evolution, identifying models which are consistent with $T_e > 300$ km and the observed recent magmatic activity. We find that although models satisfying both constraints can be constructed, special assumptions regarding the concentration and distribution of radioactive elements, the style of mantle convection and/or the mantle's volatile content need to be made. If a dry mantle rheology is assumed, strong plumes caused by, e.g., a strongly pressure dependent mantle viscosity or endothermic phase transitions near the core–mantle boundary are required to allow for decompression melting in the heads of mantle plumes. For a wet mantle, large mantle water contents of the order of 1000 ppm are required to allow for partial mantle melting. Also, for a moderate crustal enrichment of heat producing, elements the planet's bulk composition needs to be 25 and 50% sub-chondritic for dry and wet mantle rheologies, respectively. Even then, models resulting in a globally averaged elastic thicknesses of $T_e > 300$ km are difficult to reconcile with most elastic thickness estimates available for the Hesperian and Amazonian periods. It therefore seems likely that large elastic thicknesses in excess of 300 km are not representative for the bulk of the planet and that T_e possibly shows a large degree of spatial heterogeneity.

1. Introduction

In the absence of direct measurements, the elastic lithosphere thickness T_e is one of the few clues we have to the thermal state a planet, enabling us to reconstruct its thermal history. The latter significantly influences tectonic, magmatic and geological processes present on the surface and Phillips et al. (2008) have recently studied the geodynamical response of the martian lithosphere to loading by the northern polar cap to constrain the present day martian elastic lithosphere thickness.

Using radar sounding data obtained by SHARAD, the shallow radar onboard the Mars Reconnaissance Orbiter, Phillips et al. (2008) found that the martian lithosphere is extremely stiff and T_e is larger than 300 km at the north pole today. This is surprising as this value is almost twice as large as previously estimated from theoretical considerations (Grott and Breuer, 2008a). Also, the best fit elastic thickness derived from flexure studies of the south polar load yields $T_e = 140$ km, although any value greater than 102 km can fit the observations (Wieczorek, 2008).

Phillips et al. (2008) propose three solutions to explain these large elastic thicknesses: First, elastic thicknesses could be spatially heterogeneous, being small near the volcanic centers of Tharsis and Elysium and large in the polar regions. In this case, the estimated $T_e > 300$ km would not be representative for the bulk of the planet and the average elastic thickness would be somewhat smaller. Second, the Planum Boreum load could not be in dynamic equilibrium, i.e., the small amount of deflection could result from a transient state of the planetary interior. In this case, the current elastic thickness at the north pole could be smaller than 300 km. Finally, if $T_e > 300$ km is representative for the bulk of the planet, the planetary interior needs to be relatively cold and Phillips et al. (2008) propose that the bulk concentration of radiogenic elements in the martian interior could be sub-chondritic.

Of these approaches, the last two appear to be problematic: First, as shown by Phillips et al. (2008), it is difficult to construct models with small elastic thicknesses that do not result in deflections well exceeding the observed value one Myr after loading. Second, geochemical analysis of the SNC meteorites implies essentially chondritic concentrations of radioactive elements (Wänke and Dreibus, 1994; Treiman et al., 1986; Lodders and Fegley, 1997). Furthermore, if the concentration of heat producing elements in the martian interior is indeed reduced, the resulting low inte-

* Corresponding author.
E-mail address: matthias.grott@dlr.de (M. Grott).

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rior temperatures could possibly inhibit partial mantle melting and magmatism. However, geological evidence suggests that Mars has been volcanically active in the recent past (e.g., Neukum et al., 2004a).

This study will explore the implications of large elastic thicknesses and we will assume that $T_e > 300$ km is globally representative. We will reinvestigate the martian thermal evolution considering wet as well as dry mantle rheologies. Different compositional models in terms of the bulk content of heat producing elements will be investigated and models which are compatible with the large elastic thicknesses reported by Phillips et al. (2008) will be identified. Furthermore, the evolution of the elastic lithosphere thickness predicted by these models will be compared to the available elastic thickness estimates derived from gravity and topography data (McGovern et al., 2004; Belleguic et al., 2005; Hoogenboom and Smrekar, 2006; Wieczorek, 2008), forward modeling of thrust faults (Schultz and Watters, 2001; Grott et al., 2007), the analysis of rift flank uplift (Grott et al., 2005; Kronberg et al., 2007) and the new estimate by Phillips et al. (2008). Finally, we will address the question of whether these models are compatible with the production of partial melt in the martian mantle today.

2. Model

2.1. Thermal evolution and elastic thickness

The thermal evolution of Mars is modeled solving the energy balance equations for the core, mantle and lithosphere, treating the mantle energy transport by parametrized convection models using scaling laws for stagnant lid convection (Grasset and Parmentier, 1998). The model is similar to that of Grott and Breuer (2008a) and we ignore crustal production. Instead, we assume that the bulk of the crust is primordial and although there is evidence for late crustal production even after 4 Gyr (Hartmann et al., 1999; Hartmann and Berman, 2000; Neukum et al., 2004a; Grott, 2005), its volumetric contribution is probably minor on a global scale (Nimmo and Tanaka, 2005). Note that the latent heat of melting associated with crustal production can have a large influence on the total amount of melt produced during the evolution (Hauck and Phillips, 2002), but does not significantly affect the current thermal state of Mars as it is expected to change the mantle energy balance by less than one percent. Details of the model used here may be found in Grott and Breuer (2008a).

Thermal evolution models start from temperature profiles that vary adiabatically in the core and mantle and the core is assumed to be superheated with respect to the mantle by 300 K (Stevenson, 2001; Breuer and Spohn, 2003). Cooling of the planet is then controlled by mantle convection and the mantle viscosity determines the efficiency of the heat transport, the thickness of the thermal boundary layers and the temperatures in the martian lithosphere.

Given the thermal structure of the lithosphere, the elastic lithosphere thicknesses T_e is calculated using the strength envelope formalism (McNutt et al., 1988) for the two-layer system consisting of crust and mantle. Given the rheologies for the crust and mantle the elastic thicknesses of the individual crustal and mantle layers $T_{e,c}$ and $T_{e,m}$ are calculated. As the numerical results will be compared to the elastic thickness derived from the non-flexed lithosphere at the north pole, zero bending moment is assumed for the calculation of $T_{e,c}$ and $T_{e,m}$.

The elastic thickness of the compound system consisting of the crust and mantle layers then depends on whether the individual layers are welded or separated by a layer of incompetent crust. If the layers are detached, T_e is given by

$$T_e = \left(T_{e,m}^3 + T_{e,c}^3\right)^{1/3}, \tag{1}$$

where $T_{e,m}$ and $T_{e,c}$ are the thicknesses of the elastic portions of the mantle and crust, respectively (e.g., Burov and Diament, 1995). If, however, $T_{e,c}$ equals the crustal thickness and no layer of incompetent crust exists, T_e is simply given by the sum of the individual components which then act as a single plate and

$$T_e = T_{e,m} + T_{e,c}. \tag{2}$$

Details of the this model may be found in Grott and Breuer (2008a).

2.2. Recent volcanism caused by mantle plumes

There are two alternative explanations for the persistence of volcanism on Mars today. Recent production of partial melt in the martian mantle could be caused by decompression melting in the heads of uprising mantle plumes (O'Neill et al., 2007; Li and Kiefer, 2007) or by heat accumulation underneath regions of a thickened, thermally isolating crust (Schumacher and Breuer, 2007). In order to determine whether partial mantle melting is feasible on Mars today, we compare the temperature of mantle plumes to the solidus of peridotite. The effect of a thermally isolating crust will be discussed in Section 5. In our parameterized models, mantle plumes originate at the core–mantle boundary and the initial plume temperature is given by the core temperature T_c. Plumes then cool adiabatically as they rise to the surface and the plume temperature is given by

$$T_{\text{plume}} = T_c - \frac{\alpha g T_c z}{c_m}, \tag{3}$$

where α is the thermal expansion coefficient, g is the gravitational acceleration, c_m is the mantle specific heat capacity and z is the distance from the core–mantle boundary.

The solidus of peridotite is determined by laboratory experiments and we use the parameterization

$$T_{\text{sol}} = 1409 + 134.2P - 6581P^2 + 0.154P^3, \tag{4}$$

where P is the pressure in GPa (Takahashi, 1990).

Given the plume and solidus temperatures T_{plume} and T_{sol}, we compute the temperature difference ΔT_{lid} between T_{plume} and T_{sol} at the base of the stagnant lid

$$\Delta T_{\text{lid}} = T_{\text{plume,lid}} - T_{\text{sol,lid}} \tag{5}$$

to determine if partial melt is generated in the convecting mantle. Furthermore, the depth of decompression melting D_{melt}, given by the depth at which T_{plume} equals T_{sol}, is also calculated to determine how deep mantle plumes would have to penetrate into the stagnant lid to generate partial melt (compare Fig. 3).

2.3. Recent volcanism caused by hydrous mantle melting

Laboratory experiments indicate that water is important for the onset of mantle melting (Hirth and Kohlstedt, 1996; Asimow and Langmuir, 2003; Katz et al., 2003; Hirschmann, 2006) and concentrations in excess of a few hundred ppm can significantly reduce the solidus of mantle rocks. We investigate the influence of water on the solidus of peridotite using the parameterization by Katz et al. (2003). Given the bulk concentration of water in the mantle rock $X_{\text{H}_2\text{O}}^{\text{bulk}}$ and the partitioning coefficient $D_{\text{H}_2\text{O}}$ between solid and silicate melt, the concentration of water in the melt $X_{\text{H}_2\text{O}}$ can be determined from

$$X_{\text{H}_2\text{O}} = \frac{X_{\text{H}_2\text{O}}^{\text{bulk}}}{D_{\text{H}_2\text{O}} + F(1 - D_{\text{H}_2\text{O}})}, \tag{6}$$

where F is the melt fraction and $D_{\text{H}_2\text{O}} = 0.01$ (Katz et al., 2003; Aubaud et al., 2004). The maximum reduction of the peridotite