THE EVOLUTION OF THE MARTIAN LITHOSPHERE AND IMPLICATIONS FOR CRUSTAL AND MANTLE RHEOLOGY. M. Grott, D. Breuer, *Institute of Planetary Research, German Aerospace Center (DLR), Berlin, Germany (matthias.grott@dlr.de, doris.breuer@dlr.de).*

Introduction: Estimates of the Martian elastic lithosphere thickness imply that T_e evolved from values below 20 km in the Noachian to more than 100 km in the Amazonian period [1] and this general trend is well understood in terms planetary cooling as predicted by thermal evolution models [2][3]. However, the data also implies that the lithospheric thickness rapidly increased from ∼30 km to ∼70 km during the Hesperian period, which may not be explained by the depth evolution of a single isotherm. Therefore, [3] have argued that during early evolution the mechanical strength of the lithosphere is carried by the crust alone and that the elastic lithosphere thickness is determined by the depth to the onset of plasticity in diabase. Later, after the upper mantle has cooled sufficiently, it starts to contribute to the elastic strength of the lithospheric plate and T_e corresponds to the depth to the brittle-ductile transition in olivine.

Here we will expand this model to include the contributions to the elastic lithosphere thickness of both crust and mantle and use the strength envelope formalism to study the coupling between the two layers [4]. We will investigate the growth of the elastic lithosphere thickness during planetary evolution and show that the rapid increase of lithospheric thickness in the Hesperian is caused by the coupling of crust and mantle, similar to what is observed on Earths continents [4]. Furthermore, we will investigate the influence of crustal and mantle rheology and show that the evolution of the elastic lithosphere thickness is best compatible with wet rheologies for both crust and mantle.

Elastic Thickness Data: We have compiled estimates of the Martian elastic lithosphere thickness T_e which have been derived from gravity and topography data [1][5], forward modeling of thrust faults [6][7] and the analysis of rift flank uplift [8][9]. The data of [1] and [5], which give estimates for the same structures and loading events, have been merged such that the estimates are compatible with both studies. Furthermore, only those data-points of [1] have been used, which give upper as well as lower bounds for T_e.

The resulting dataset is shown in Fig. 1. Data-points have been merged into time-bins corresponding to the Noachian, Noachian-Hesperian, Hesperian, Hesperian-Amazonian and Amazonian epochs. The time coordinate for each individual data-point has been chosen such that the points plot around the center of their respective time periods. Mean elastic thicknesses as well as the spread of the data have thus been calculated. They are indicated by the shaded rectangles in Fig. 1. Two outlier data-points have been disregarded for the averaging process and are indicated by circular symbols.

During the Noachian to Hesperian periods, observed elastic thickness values range from ∼10 km at rifts [8][9] to about 30 km at lobate scarps [6][7]. Average T_e ranges from 21±9 to 25±7 and 30±9 km during the Noachian, Noachian-Hesperian and Hesperian periods, respectively. Then, T_e rapidly increases to 62±30 and 68±36 km during the Amazonian-Hesperian and Amazonian periods, respectively. In the following we will use this elastic thickness history, in particular the rapid lithospheric growth in the Hesperian period, to constrain the crustal and mantle rheology.

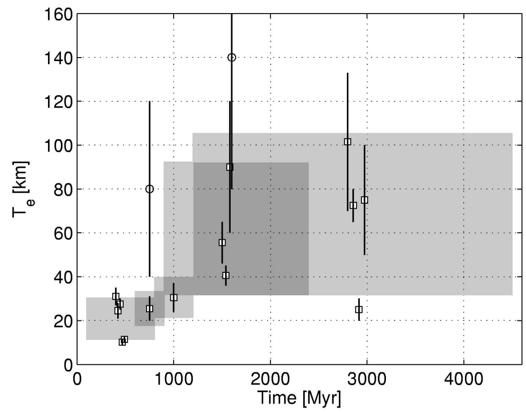

Figure 1: Compilation of elastic thickness estimates as a function of time. Data-points have been merged into time-bins gray rectangles are centered around the mean elastic thickness of the corresponding epoch, their heights correspond to one standard deviation.

Thermal Evolution Model: We calculate the thermal evolution of Mars by solving the energy balance equations for the core and mantle, treating the mantle energy transport by parametrized convection models. This is done using scaling laws for stagnant lid convection [10] and our model is similar to that of [3]. We ignore crustal production and assume that the bulk of the crust is primordial. Although there is evidence for late crustal production even after 4 Gyr [11], the volumetric contributions are probably minor on a global scale [12].

Models of the thermal evolution of Mars need to satisfy the constrains given by the observed average crustal thickness [13] and the magnetic field history. [14] found that both wet and dry mantle rheologies are compatible with these constrains if initial mantle temperatures of 1700 to 2000 K are assumed. Furthermore, the magnetic field history requires either an early episode of plate tectonics or a superheated core [15][16]. The nominal model of [2] tends towards initial mantle temperatures of 1700-1800 K and a wet mantle rheology.

More recently, the influence of a brecciated upper crust and the temperature dependence of thermal conductivity on the thermal evolution have been considered [3]. The presence of a low conductivity crust offers an explanation for the ongoing volcanism in selected provinces like Tharsis. However, thermal insulation would also increase the temperatures

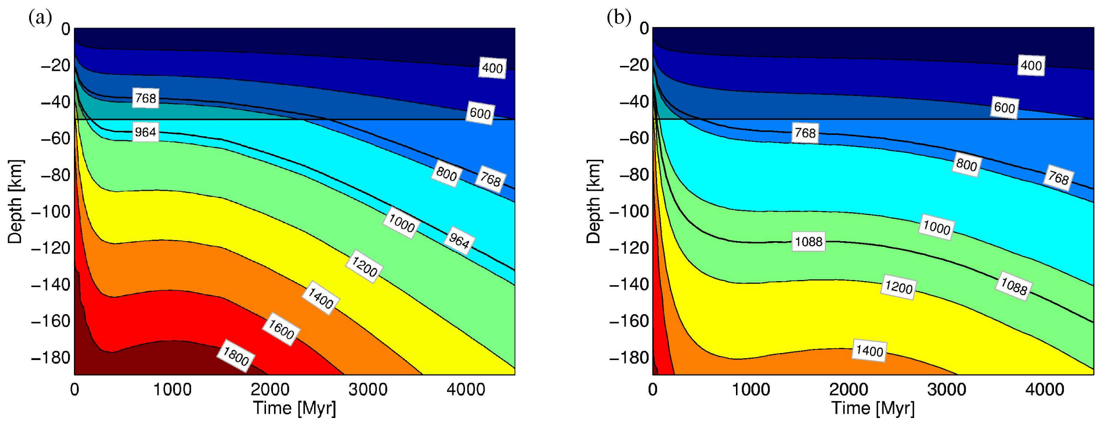

Figure 2: Contour plots of the temperature T as a function of time and depth for the reference thermal evolution model. The isotherms corresponding to the loss of mechanical strength in wet diabase (768 K), wet olivine (964 K) and dry olivine (1088 K) are also indicated. The base of the crust is marked by the horizontal line and the area above is slightly shaded. (a) Evolution assuming a wet mantle and $\eta_0 = 10^{20}$ Pa s. (b) Evolution assuming a dry mantle and $\eta_0 = 10^{21}$ Pa s.

at the crust-mantle boundary, initiating lower crustal flow and relaxing surface topography, contrary to observations. The insulating effects of a brecciated crust may however be reduced by hydrothermal circulation in the upper crust, thus increasing its overall conductivity [17].

Our nominal model has the following parameters: The mean crustal thickness D_c is taken to be 50 km, consistent with gravity and topography data [13]. The initial upper mantle temperature is 1800 K, the crustal thermal conductivity k_c is 2.5 W m^{-1} K^{-1} and the enrichment of radioactive elements in the crust relative to the undepleted mantle Λ is a factor 5. Depending on the rheology of the mantle, the reference viscosity η_0 at 1600 K is chosen to be 10^{20} and 10^{21} Pa s for a wet and dry mantle, respectively. The thermal evolution of these models in shown in Fig. 2, where contour plots of the temperature T are shown as a function of time and depth for the wet (a) and dry (b) mantle. The crustal thickness is indicated by the horizontal line and the area above the line is slightly shaded. The isotherms corresponding to the loss of mechanical strength in wet diabase (768 K), wet olivine (964 K) and dry olivine (1088 K) are also indicated.

For a wet mantle rheology, mantle convection is more vigorous and the thermal boundary layers at the top and bottom of the mantle are very thin. Thus the planet may efficiently cool and a lot of heat is transferred into the base of the lithosphere, which then becomes thin and hot. Therefore, the elastic portion of the lithosphere is much smaller for the wet than for the dry mantle. As long as the lower crust is sufficiently hot, a mechanically incompetent layer decouples the crust from the mantle and the lithosphere behaves as a multilayer elastic plate. Once the lower crust cools below 768 K and becomes mechanically strong, the lithosphere behaves as a single plate. For the wet and dry mantle, this transition occurs at 2700 and 500 Myr, respectively, and results in a rapid increase of the elastic lithosphere thickness.

Strength Envelopes: Given the thermal structure of the Martian lithosphere, we calculate the elastic thickness assuming a two-layer system consisting of crust and mantle with corresponding rheological properties. The elastic thickness of the individual layers is calculated from strength envelopes, assuming a non-flexed plate, i.e. zero bending moment. In this case the mechanical thickness equals the elastic thickness and T_e is limited by the depth at which the plate looses its mechanical strength due to ductile flow. Assuming a bounding stress of $\sigma_y = 15$ MPa [4], the corresponding temperature and thus depth may be calculated from

$$T(\sigma_y) = \frac{Q}{R}\left[\log\left(\frac{\sigma_y^n A}{\dot{\epsilon}}\right)\right]^{-1} \tag{1}$$

where A, n and Q are rheological parameters derived from laboratory experiments, R is the gas constant and $\dot{\epsilon}$ is strain rate. The different rheologies used in this study and the resulting limiting temperatures are summarized in Table 1

The elastic thickness of the compound system then depends on whether the individual layers are welded or separated by a layer of incompetent crust. Then, T_e of the detached plate is calculated from

$$T_e = (T_{e,m}^3 + T_{e,c}^3)^{\frac{1}{3}} \tag{2}$$

where $T_{e,m}$ and $T_{e,c}$ are the thicknesses of the elastic portions of the mantle and crust, respectively (e.g., [4]). If, however, $T_{e,c}$ equals the crustal thickness, T_e is simply the sum of the individual components which then act as a single plate

$$T_e = T_{e,m} + T_{e,c} \tag{3}$$

Fig. 3 shows a representative strength envelope for our reference model and a wet mantle rheology. Coupling of crust and mantle occurs once the incompetent lower crust has vanished, which happens around $\sim$2700 Myr .

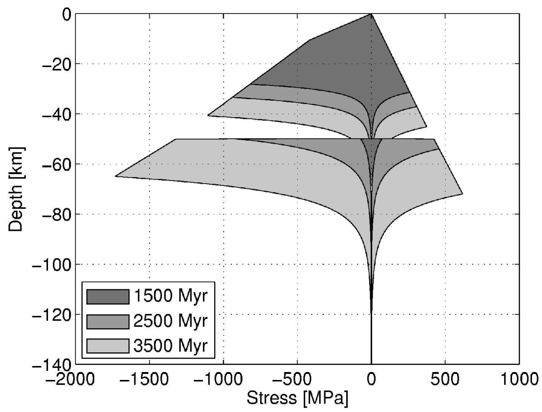

Figure 3: Yield strength envelopes for the reference model at different times assuming a wet diabase crust and wet olivine mantle rheology.

Results: We have investigated the influence of crustal and mantle rheology on the evolution of the lithospheric thickness and the results are summarized in Fig. 4. As a reference, a simplified representation of the observed elastic thickness values (Fig. 1) is given by the shaded rectangles.

For a wet mantle and crustal rheology (Fig. 4a), rapid growth of the lithospheric thickness is observed around 2200 Myr, depending on initial upper mantle temperature T_m. The magnitude of the increase is $\sim$35 km and in good agreement with the observations. Although the jump occurs fairly late, the exact timing was found to also depend on other model parameters, the most important of which turned out to be the initial crustal thickness. For $D_c = 40$ km, the jump occurs at 1200 Myr, in very good agreement with the observations (not shown). The general trend of T_e evolution is satisfactorily reproduced, with values around 30 km in the early and $\sim$100 km during the late evolution.

If the mantle is assumed to be dry (Fig. 4b), mantle convection becomes less vigorous and the thickness of the thermal boundary layers increase. As a consequence the stagnant lid becomes much cooler and lithospheric strength is significantly increased. Coupling of crust and mantle will then typically occur much earlier, except for initially very hot models. Additionally, the mechanical strength of the mantle is also increased which then results in large T_e very early in the evolution, which is difficult to reconcile with the observations.

If the mantle is significantly weaker than the crust, i.e. for a wet mantle and dry crust (Fig. 4c), lithospheric growth is continuous. The layer of incompetent lower crust vanishes before the mantle obtains significant strength and the lithosphere therefore acts as a single plate during the entire evolution. This results in large elastic thickness values even for early times, contrary to the observations.

Finally, if both mantle and crust are assumed to be dry (Fig. 4d), the two-layer rheology again acts like a single plate and mechanical strength is significantly increased with respect to wet rheologies. Furthermore, the lithosphere itself is much cooler, again adding to the elastic thickness. Therefore, T_e reaches very high values early on in the evolution even for high initial mantle temperatures, which is incompatible with the observations.

Discussion: Elastic thickness data indicate that the Martian elastic lithosphere was initially relatively thin and increased its thickness rapidly during the Hesperian period. We have modelled the thermal evolution of Mars and studied the influence of crustal and mantle rheology on the evolution of the lithospheric thickness and found that the observations are best compatible with a wet crust and wet mantle rheology. A dry Martian crust can be ruled out as these models fail to reproduce the rapid lithospheric growth and low elastic thickness values during the Noachian and early Hesperian periods.

So far, we have assumed that no bending stresses are acting on the lithospheric plate. However, flexure due to loading will reduce T_e by brittle and ductile yielding in the top and bottom layers, respectively. Furthermore, the associated bending stresses could cause a decoupling of the crust from the mantle, further reducing the elastic thickness of the entire plate. This would make it easier to reconcile the results for the dry mantle with the observations, such that a dry mantle rheology cannot be ruled out at present. However, if the Martian mantle were dry, high initial upper mantle temperatures would be required to yield the observed low elastic thickness values during the Noachian.

The presented results are consistent with an early, deep seated global scale hydrological system [17][18], which would result in a wet crustal rheology. Furthermore, a wet crust could provide the pore fluid necessary for hydrothermal crustal cooling, thus preventing viscous relaxation of topography during the early Martian evolution [17].

References: [1] McGovern et al. (2004), *J. Geophys. Res.*, 109, E07007. [2] Hauck and Phillips (2002), *J. Geophys. Res.*, 107, E7, 5052. [3] Schumacher and Breuer(2006), *J. Geophys. Res.*, 111, E02006. [4] Burov and Diament (1995), *J. Geophys. Res.*, 100, B3, 3905-3927. [5] Belleguic et al. (2005), *J. Geophys. Res.*,110, E11005. [6] Schultz and Watters (2001), *Geophys. Res. Lett.*, 28, 24, 4659-4662. [7] Grott et al. (2007), *Icarus*, 186, 517-526. [8] Grott et al. (2005), *Geophys. Res. Lett.*, 32, L21201. [9] Kronberg et al. (2007),*J. Geophys. Res.*, in press. [10] Grasset and Parmentier (1998), *J. Geophys. Res.*,103, 18171-18181. [11] Grott (2005), *Geophys. Res. Lett.*, 32, L23201. [12] Nimmo and Tanaka (2005), *Annu. Rev. Earth Planet. Sci.*, 33, 133-161. [13] Neumann et al. (2004), *J. Geophys. Res.*, 109, E08002. [14] Breuer and Spohn (2006), *Planet. Space Sci.*, 54, 153-169. [15] Breuer and Spohn (2003), *J. Geophys. Res.*, 108, E7, 5072. [16] Nimmo and Stevenson (2000), *J. Geophys. Res.*, 105, E5, 11969,11979. [17] Parmentier and Zuber (2007), *J. Geophys. Res.*, 112, E02007. [18] Andrews-Hanna et al. (2007), *Nature*, 446, 163-165. [19] Russel and Head (2007), *Planet. Space Sci.*, 55, 315-332. [20] Mackwell et al. (1998), *J. Geophys. Res.*, 103, B1, 975-984. [21] Caristan (1982), *J. Geophys. Res.*, 87, 6781-6790. [22] Karato et al. (1986), *J. Geophys. Res.*, 91, 8151-8176.

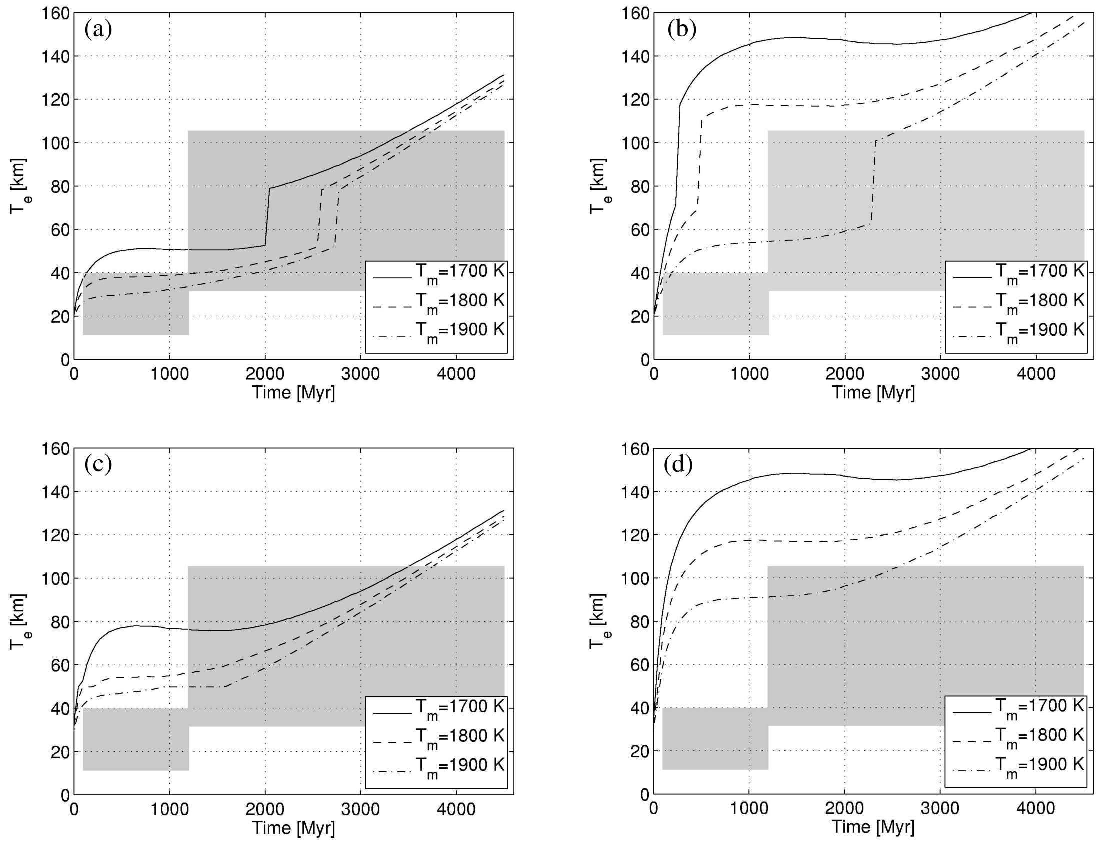

Figure 4: The elastic thickness T_e as a function of time, varying initial upper mantle temperature as well as crustal and mantle rheology. For the wet and dry mantle we have assumed reference viscosities η_0 of 10^{20} and 10^{21} Pa s, respectively. (a) Wet mantle, wet crust, (b) Dry mantle, wet crust, (c) Wet mantle, dry crust, (d) Dry mantle, dry crust . The parameter ranges corresponding to the observed elastic thickness values are indicated by the shaded rectangles.

Table 1: Rheological parameters used in this study.

Rheology	A [Pa^{-n} s^{-1}]	n	Q [kJ mol^{-1}]	$T(\sigma_y)$ [K]	Reference
Diabase(dry)	1.1×10^{-26}	4.7	488	1047	[20]
Diabase(wet)	3.1×10^{-20}	3.05	276	768	[21]
Olivine(dry)	2.4×10^{-16}	3.5	540	1088	[22]
Olivine(wet)	1.9×10^{-15}	3.0	420	946	[22]