THERMAL AND VOLATILE EVOLUTION OF MARS FROM DICHOTOMY RELAXATION MODELS.

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Introduction: The Noachian epoch on Mars included the formation of the global dichotomy boundary between the northern plains and southern highlands [1,2], the majority of the crust [1 & references therein], and phyllosilicates [3,4]. This early time period was one of high interior heat flow and relatively wet conditions at least at the surface. Given these initial conditions, the dichotomy boundary should have relaxed more fully for most plausible assumptions. In this study we show that loss of crustal volatiles at the Hesperian/Amazonian boundary is needed along with cooling predicted by thermal evolution models to preserve the dichotomy topography and crustal thickness variations to the present day. For plausible parameter ranges, the effect of interior volatiles has a stronger effect on relaxation than mantle temperature and interior processes such as mobile vs. stagnant lid convection or global mantle overturn.

Approach: In this study we predict relaxation of the dichotomy using present day dichotomy topography and estimates of elastic thickness over time as constraints. We model the relaxation as function of evolution of the temperature and stress over the first 1 Ga after formation of the dichotomy, when the majority of relaxation occurs. Temperature change is based on three different interior evolution models: 1) stagnant lid convection [5], 2) plate tectonics until 4 Ga followed by stagnant lid convection [6,7], and 3) mantle overturn caused by unstable density stratification created in a magma ocean [8,9]. We recreate the thermal evolution predicted by each of these models by solving the heat conduction equation in the lithosphere using initial and boundary conditions based on these three models. The thermal evolution calculation simulates mantle cooling or heating (for the stagnant lid model), and includes radiogenic heat production.

The stagnant lid model starts with the coolest lithosphere, but does not cool as efficiently as the other two models (see Fig. 1). The early plate tectonics and mantle overturn models cool with similar efficiency, though the mantle overturn model creates a thicker thermal lithosphere (Fig. 1).

Model: We use a semi–analytic topographic relaxation model [10] that represents the topographic and crust-mantle surfaces as sin and cosine functions. Model geometry and standard parameters are shown in Figure 2. This model is based on the relaxation of a density-stratified incompressible fluid [11,12] in which horizontal variations of stress and displacement are transformed to the frequency domain and vertical variations in viscosity are numerically integrated over depth. Two density layers (crust and mantle) and three viscosity layers (upper crust, lower crust, and mantle) are included. The constitutive and equilibrium equations are solved for a series of time steps. Viscoelastic deformation is modeled by applying a Laplace transform to the viscous constitutive equation, giving an equation with the same form as that for a viscous in compressible fluid. Viscosity is updated at each time step. Since strain rate is dependent on temperature and stress, viscosity varies continuously with depth and time. The initial stress distribution is defined by prior finite element modeling of dichotomy relaxation [10], and varies from 2 MPa at the surface to 20 MPa at the base of the crust. This distribution is approximated with 3 layers, which provides a good approximation to finite element model calculations [10]. The viscosity of the upper crust, or elastic layer, is 10^{29} Pa s. The elastic thickness is based on the initial thermal gradient and increases with time, in some cases migrating into the mantle layer. The initial mantle viscosity for the stagnant lid and early plate tectonics models is the same; for mantle overturn the value is an order of magnitude higher. As stress in the mantle remains essentially constant, viscosity variations are a function of temperature.

The time evolution of the stress in the lower crust is difficult to predict. For strong rheologies, we assume no zero change in the strain rate and an exponential decay of stress with time. For weak rheologies, we assume strong lower crustal flow, giving a viscosity similar to the mantle. Note that although there are uncertainties in modeling evolution of the viscosity with time in this manner, the approach has been benchmarked against finite element models [10] and provides a much more accurate model of the rheology predicted by stress and thermal evolution with time than models that assume a constant viscosity.

We examine the effects of plausible ranges for the key factors controlling the rate of dichotomy relaxation: thickness of the crust, the volatile content of the crust and mantle, initial mantle temperature, and dichotomy height and slope. Standard model parameters are: 62 km crustal thickness, 1700 K mantle temperature, and 220 K surface temperature. We also consider a hotter mantle temperature (1900 K) and thinner crust. The nominal model of Hauck and Phillips [5] predicts that 75% of the 62 km thick crust is formed by 4 Ga.

Using this average value of crustal thickness (or 47 km in the plains and 77 km in the highlands) allows for comparisons between models and is consistent both with the other interior evolution models and constraints from gravity and topography data [13,14]. We consider wet and dry rheologies for the crust and mantle [15-17].

Constraints from Elastic Thickness Estimates: Elastic thicknesses predicted in our models are compared with published values determined from modeling of the gravity and topography data [18-20]. The surface ages of regions modeled range from Noachian to Amazonian. Although there are certainly local effects coming into play, as well as model uncertainties, overall the variations from small elastic thickness values (0-30 km) in the Noachian to larger values (60 to >120 km) for Amazonian age terrains are believed to reflect global planetary cooling [18].

Constraints from Topographic Relaxation: We distinguish 3 types of predicted relaxation: relaxed, partially relaxed, and unrelaxed. Only 3 types are considered due to both model uncertainties and unknowns in the initial height and geometry of the dichotomy. Figure 3 shows the comparison of the initial model topography, predicted relaxation over 1 Ga, and averaged MOLA topography in a relatively unmodified region of the dichotomy. Detailed models have shown that the dichotomy boundary is partially relaxed [10,21,22], thus we consider models that allow partial relaxation to be successful. Successful models have a wet crust and mantle and cold mantle temperatures. Models with dry crustal rheology are unrelaxed. Models with a high mantle temperature allow too much relaxation. Essentially the viscosity of the lower crust has to be 10^{20}-10^{21} Pa s in the first 0.1 Ga after formation of the dichotomy to allow short wavelength relaxation but preservation of the longer wavelength topography. Any of the three thermal models can produce this viscosity range for either a hot, dry crust, or a wet, cold crust.

Results: For the stagnant lid model thermal evolution model (Figure 4a), weaker, wetter rheologies are not consistent with relatively low elastic thickness values in the Noachian, while drier, stronger models are needed to fit higher elastic thicknesses in the Hesperian/Amazonian. The result is similar for the early plate tectonics model (Fig. 4b), though thicker lithosphere shifts elastic thickness to earlier times. For the mantle overturn model (Fig. 4c), which implies formation of dichotomy very early, perhaps at ~4.4 Ga, elastic thicknesses are larger at earlier times.

Discussion: Additional information on interior volatile content must also be considered. Analysis of meteorite data suggest that water was present in the crust or mantle in at least one location on Mars at 175 Ma [23], though other areas suggested a source depleted in water [24]. Meteorite data also indicate very early formation of the crust [1]. A wet, low viscosity mantle is needed to do this for a stagnant lid regime [5]. However the same is not true mantle overturn.

From this study, and other studies of dichotomy relaxation [10, 21], partial relaxation of the dichotomy is most sensitive to lower crustal viscosity, implying either a hot (~2000K), dry crust, or a cold (~1500K), wet crust. The later case is consistent with estimates from the thermal models used here for initial temperatures, as well as general evidence for a wet, early Mars. Wet crust and mantle are also consistent with elastic thickness estimates from the Noachian (Figure 4).

The prediction of elastic thickness assuming wet crust and mantle begins to break down either in the Noachian-Hesperian transition or in the Hesperian, depending on which interior model is assumed and whether or not the elastic thickness estimate from S. Hellas w/intrusion is considered an outlier. If we disregard that particular elastic thickness value, the stagnant lid and early plate tectonic models with wet conditions no longer agree in the Hesperian (Figure 4). The mantle overturn model predicts too large an elastic thickness value in the Noachian-Hesperian time frame (Figure 4). These models only agree with elastic thickness values if there is a loss of volatile from at least the crust. The mantle could be either dry or wet.

The local nature of elastic thickness estimates must be considered. The highlands may differ from the north in their thermal evolution, and the Tharsis area is clearly unique. In particular, extended volcanism at Tharsis may have acted to dry the crust and/or mantle, yielding the much larger elastic thickness values.

Dichotomy relaxation models agree well with other evidence for wet rheology early in Martian history. The comparison of elastic thicknesses predictions from these models to estimates from admittance implies that a loss of crustal volatiles is needed, at least in areas of high volcanic activity. Note that such areas provide the only available estimates of elastic thickness from the Hesperian. This result may be consistent with the observation that phyllosilicates occur only in the Noachian [3,4,25]. Volcanic activity may have provided the water necessary to form the observed phyllosilicates.

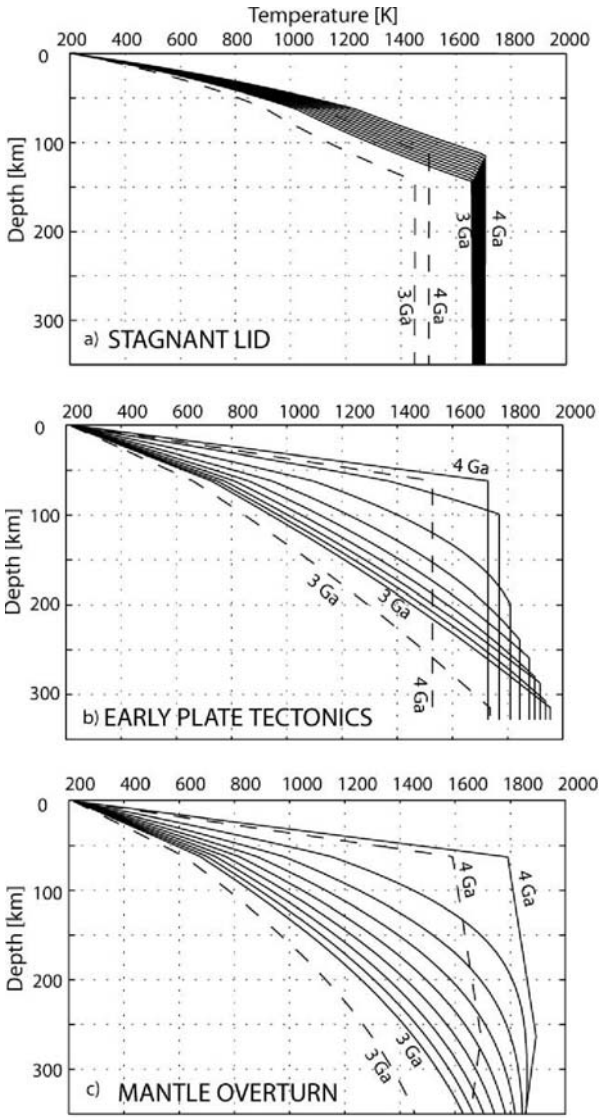

Figure 1: Thermal evolution for three interior evolution models of the lithosphere from 4 to 3 Ga. The dashed lines show a variant of each model in which the mantle temperature is 200 K cooler than that in the nominal model.

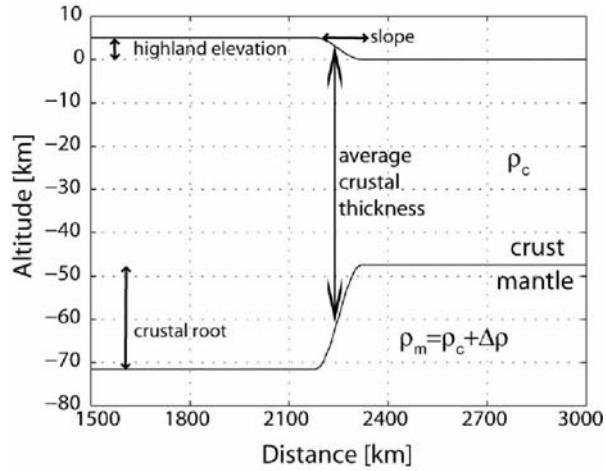

Figure 2: Sketch of the semi-analytical model to the depth of 80 km. Horizontal size of the model is 4500 km, but only the central third is shown. The initial topography is slightly higher and steeper than present-day dichotomy topography. The topography is isostatically compensated (Airy model) by a crustal root (density is 2900 kg/m^3) at the crust/mantle boundary (the density difference is 600 kg/m^3).

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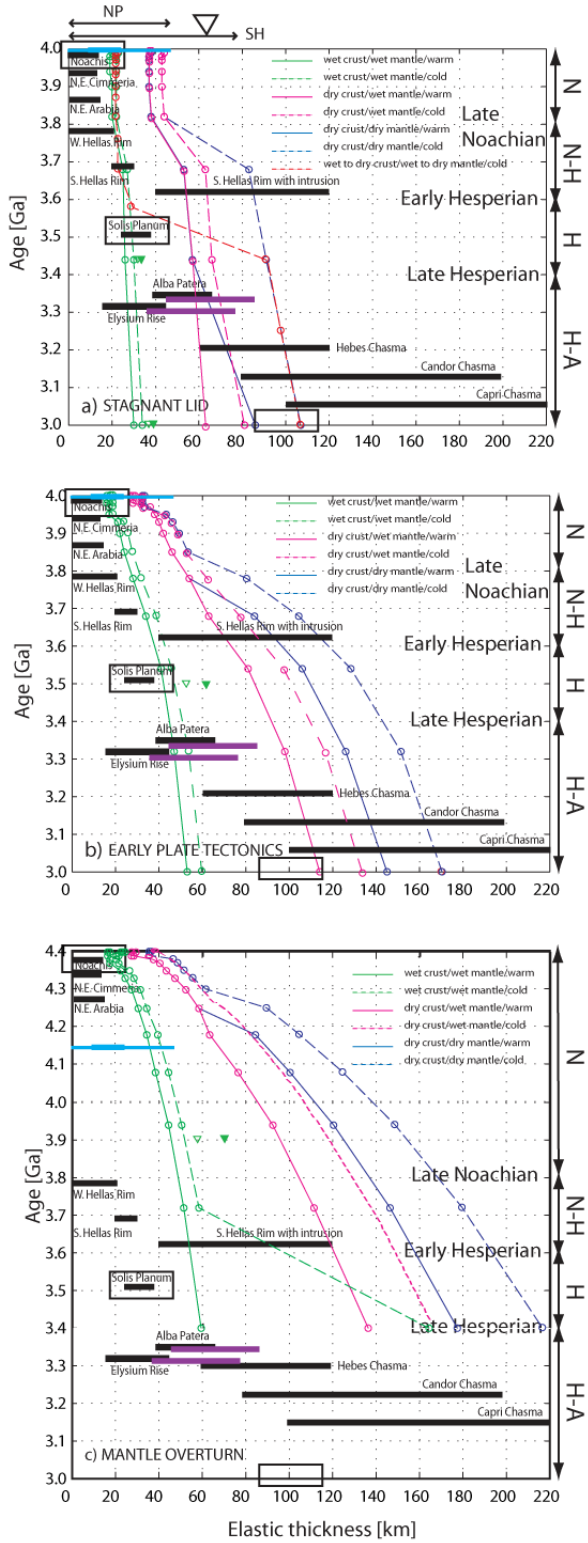

Figure 3: Comparison between modeled and admittance elastic thicknesses for a) stagnant lid, b) early plate tectonics, and c) mantle overturn thermal models. The modeled elastic thickness is shown for a range of wet and dry crust and mantle rheologies. When wet crustal rheology is used, two elastic layers develop; then green lines show the elastic thickness in the crust, while magenta lines deeper than base of the crust (62 km) show the elastic thickness in the wet mantle. The effective thickness, shown by empty (cold thermal model) and full (warm thermal model) triangles, is determined using [26]. Elastic thickness estimates from admittance modeling are shown in black [18.19], grey [20], and purple [21]. The left vertical axis corresponds to the age of the modeled elastic thickness whereas the right vertical axis shows the surface age of the admittance elastic thickness (N: Noachian, H: Hesperian, A: Amazonian epochs, with divisions based on [27]. The black rectangles show the elastic thicknesses that are well constrained in time. The black double-headed arrow on top of the figures shows the crustal thickness of the northern plains ("NP") and southern highlands ("SH"). The triangle shows the average crustal thickness of 62 km.

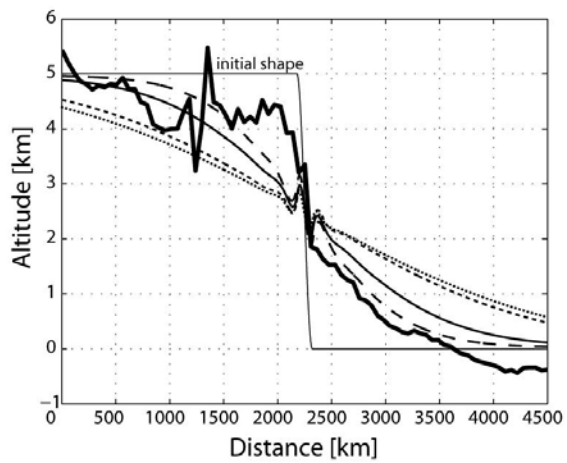

Figure 4: Results from semi-analytical relaxation modeling. The initial shape at 4 Ga, the relaxed shape at 3 Ga (thin lines), and present MOLA topography profile through the dichotomy boundary (thick line) are compared. The models with 62 km thick crust, cold mantle and initial wet crustal and mantle rheology are shown: solid line – stagnant lid model; long dashes – the same, but rheology changes to dry at 3.5 Ga; dots – early plate tectonics model; short dashes – mantle overturn model. MOLA topographic profile is averaged over 10 degree wide swath centered at 30 N and 60 E and perpendicular to the dichotomy boundary.