MAXIMUM MARTIAN CRUSTAL THICKNESS FROM VISCOUS RELAXATION OF TOPOGRAPHY. F. Nimmo, *Bullard Labs, Madingley Road, Cambridge, CB3 0EZ, UK, (nimmo@esc.cam.ac.uk)*, D. Stevenson, *Division of Geological and Planetary Sciences, California Inst. Technology, Pasadena, CA91125, USA, (djs@gps.caltech.edu).*

Topographic contrasts at the surface of a planet result in lateral pressure gradients at depth. On Earth, the lower crust in some areas is weak enough to flow in response to these gradients, reducing the topography. The rate of flow increases with increasing crustal thickness or decreasing viscosity. Since there are large topographic contrasts on Mars which have survived for 4 Ga, we can estimate the maximum crustal thickness, given an assumed rheology and heat flux. This upper bound is about 60 km.

Mars shows topographic variations at different length scales. The boundary between the southern and northern hemispheres shows an increase in elevation of 2-4 km over several hundred km laterally. The largest impact basins are 1000-2000 km in diameter and probably formed prior to 4 Ga b.p. The younger northern volcanic rises are up to 500 km across. All such topographic contrasts may be reduced by lower crustal flow.

The characteristic time τ for viscous relaxation of topography due to flow in a layer in which viscosity η increases exponentially upwards with a lengthscale δ according to

$$\eta = \eta_0 \exp(z/\delta) \tag{1}$$

is given by Ojakangas & Stevenson (1989):

$$\tau = \frac{\eta_0}{2g\Delta\rho k^2 \delta^3} \tag{2}$$

where η_0 is the viscosity at the base of the layer, g is gravity, $\Delta\rho$ is the density contrast between the layer and the underlying material and k is the wavenumber of the topography.

The effective viscosity of real materials is governed mainly by temperature, strain rate and three rheological constants. The temperature at any depth in the crust is governed by the mantle heat flux and the abundance of crustal radiogenic materials. For a given temperature profile and rheology, the variation in viscosity as a function of depth near the base of the crust may be approximated by equation (2) where

$$\delta = \frac{nRKT_b^2}{QF_b}. \tag{3}$$

Here Q and n are the activation energy and exponent of the material, R is the gas constant, K is the thermal conductivity, and F_b and T_b are the heat flux and temperature, respectively, at the base of the crust.

In order to calculate the Martian heat flux at 4 Ga we assume radiogenic concentrations identical to those of Sun & McDonough (1989). We calculate the temperature profile by assuming that a crust of thickness t_c is enriched in radiongenic elements by a factor p and that the remainder reside in the mantle. The relaxation time for a given crustal thickness can then be calculated using equations (2) and (3).

In Figure 1, we plot t_1, the crustal thickness for which the relaxation time is 1 Ga, as a function of wavelength for three different rheologies. Since this time is one quarter of the likely age of most of the Martian crust, t_1 may be regarded as an upper bound on crustal thickness. The mantle heat flux (prior to crustal extraction) is 67 mW m^{-2}, appropriate for Mars at 4 Ga, and the crustal enrichment factor p is 4. The effect of changing p to 1 is shown by the dotted line to be minor. For wavelengths of 1000-2000 km, similar to the large impact basins in the southern hemisphere of Mars, the maximum crustal thickness is less than 60 km, even for dry diabase. Shorter wavelength features would relax more rapidly, although elastic support becomes more important at shorter wavelengths. Since Mars currently possesses large amounts of ice, a weaker rheology such as that of diabase is probably more appropriate.

Figure 1 assumed a strain rate of $10^{-15}\ s^{-1}$. Varying this value by an order of magnitude changes t_1 by less than 10 km. Varying the value of p was shown above to have an even smaller effect. The largest uncertainty comes from the estimated concentration of radiogenic elements in the Martian mantle. Model concentrations vary by about $\pm$ 20 %. This uncertainty results in a variation in t_1 of $\pm$ 10 km. Thus, an extreme upper bound on the maximum likely southern hemispheric crustal thickness at 4 Ga is 80 km; a more likely value is 50-60 km.

Isostatic compensation suggests that the crustal thickness of the northern hemisphere is about 20 km less than that of the south, resulting in an upper bound on the average Martian crustal thickness of 40-50 km. This estimate is compatible with other estimates of Martian crustal thickness derived from gravitational or geochemical constraints. A Martian crustal thickness of 40 km could be produced by a spreading centre with a potential temperature of 1400°C (McKenzie & Bickle 1988).

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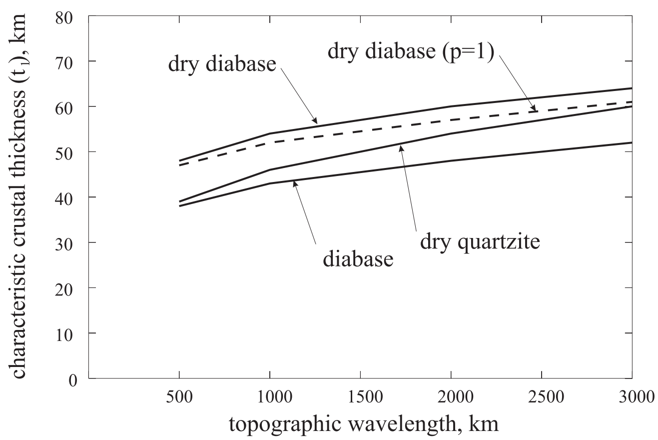

Figure 1: Characteristic crustal thickness t_1 (see text) for Mars at 4 Ga b.p. as a function of wavelength and rheology. Undepleted mantle heat flux 67 mW m^{-2}, strain rate 10^{-15} s^{-1}. Solid lines are for crustal enrichment factor p=4; dashed line for $p = 1$. Rheologies for diabase, dry diabase and quartzite are given by Shelton & Tullis (1981), Mackwell et al. (1995) and Turcotte & Schubert (1982), respectively.