Constraining the crustal thickness on Mercury from viscous topographic relaxation

Francis Nimmo

Department of Geological Sciences, University College London, England

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1. Introduction

[2] The topography of Mercury is poorly known, with only limited radar and stereo coverage available [Cook and Robinson, 2000; Harmon and Campbell, 1988]. However, radar profiles reveal topographic contrasts of several kilometers over wavelengths of ~1000 km [Harmon and Campbell, 1988]. The bulk of Mercury’s geologic activity took place within the first 1 Ga of the planet’s history [Spudis and Guest, 1988], and it is therefore likely that these topographic features derive from this period. On Earth, long wavelength topographic features are supported either convectively, as at Hawaii, or through some combination of isostasy and flexure, e.g. Tibet. Current images of Mercury show no evidence for either plate tectonics or plume activity [Melosh and McKinnon, 1988]; it was therefore assumed that neither convective support nor Pratt isostasy are operating.

[3] The composition and structure of the crust of Mercury are almost unknown. The reflectance spectrum of the surface of Mercury is similar to that of the lunar highlands [Vlas, 1988; Sprague et al., 1997], which are predominantly plagioclase. The radar characteristics of the surface are also reminiscent of the lunar highlands [Harmon, 1997]. Colour image data suggest volcanic, and possibly anorthositic, surface materials [Robinson and Lucey, 1997]. The mean density of the planet implies a thickness of mantle plus crust of around 600 km [Schubert et al., 1988]. Anderson et al. [1996] used the observed centre-of-mass centre-of-figure offset together with an assumption of Airy isostasy to infer a crustal thickness of 100–300 km. Based on tidal despinning arguments, the early elastic thickness of the lithosphere was ≤~100 km [Melosh, 1977].

[4] The temperature structure of Mercury at 4 Ga B.P. is also poorly constrained. Because of its proximity to the Sun and the effects of large impacts, Mercury may well be depleted in volatile elements such as potassium [Lewis, 1988] which contribute to radioactive heating. However, abundances of these elements are currently dependent on models resulting in uncertainties of at least an order of magnitude [Goettel, 1988; Lewis, 1988].

[5] This paper will argue that one bound on the early temperature structure of Mercury is provided by observations of ancient thrust faults. On Mercury, thrust faults with lengths of up to 500 km with a probable age of about 4 Ga are known to exist [Watters et al., in press; Watters et al., 1998; Melosh and McKinnon, 1988]. Modelling of the surface deformation associated with these faults implies fault depths of 35–40 km [Watters et al., in press].

[6] On Earth, the maximum depths of intraplate earthquakes appear to be thermally controlled, with the relevant isotherm being about 500–700 K for continental crust [Chen and Molnar, 1983] and 1000–1100 K for oceanic mantle [Wiens and Stein, 1983]. The difference between the two is probably due to the stiffer rheology and greater dryness of the oceanic material [Maggi et al., 2000]. Because the crust of Mercury is probably dry, and thus rigid, the temperature defining the base of faulting is likely to be higher than in similar material on Earth. The temperature defining the base of the faults on Mercury is therefore probably ~700 K or more. The inferred fault depth thus implies that the heat flux at 4 Ga B.P. was probably greater than 20 mW m−2 for a linear temperature gradient.

[7] On Earth, elastic thickness and seismogenic thickness are correlated, with the latter generally being slightly greater [ McKenzie and Fairhead, 1997]. For an elastic thickness of 40 km, topography at 1000 km wavelength is likely to be about 70% compensated [Turcotte and Schubert, 1982]. Large impact structures on Mercury show subdued topography and are inferred to be in a state close to total isostatic compensation [Schaber et al., 1977]. Isostatic compensation generally requires lateral variations in crustal thickness; as argued below, such variations will lead to lower crustal flow and topographic relaxation unless the crust is thin or cold.

2. Theory

[8] If topography is supported by variations in crustal thickness, pressure gradients exist which may cause the lower crust to flow, thus reducing the topography. The timescale over which this flow occurs depends on the temperature at the base of the crust, the thickness of the crust, and the composition of the crustal material.

[9] For the case in which heating is predominantly from below, an expression for the characteristic relaxation time due to lower crustal flow was obtained by Nimmo and Stevenson [2001] (hereafter NS01). Their approximations, however, break down if the heating occurs entirely from within the crust. In the latter case, the temperature T(z) is given by

\[
T(z) = T_s + \frac{H}{2K} (z^2 - z_0^2)
\]
where $H$ is the volumetric internal heat generation rate, $z$ is distance (measured upwards) from the base of the crust, $D$ is the crustal thickness, $K$ is the thermal conductivity and $T_s$ is the surface temperature.

[10] Materials deform at a strain rate $\dot{\varepsilon}$ which is strongly temperature-dependent and may also depend on the stress applied. For the temperature structure of equation (1) the stress-strain rate relationship may be written

$$\dot{\varepsilon} = \frac{\partial u}{\partial z} = A|\tau|^n \exp\left(-\frac{Q}{RT_s}\right) \sim B e^{-\frac{\varepsilon}{T_s}}$$

(2)

where $B$ is a term incorporating all variables except the depth-dependence. Here $A$ and $n$ are material constants, $\tau$ is the stress, $u$ is the horizontal velocity, and $Q$ and $R$ are the activation energy and gas constant, respectively. The characteristic channel thickness $\varepsilon$ is given by

$$\varepsilon = T_{bs} \left(\frac{2KR}{dD}\right)^{1/2}$$

(3)

where $T_{bs}$ is the temperature at the base of the crust.

[11] For lower crustal flow driven by isostatically compensated topographic contrasts, the driving pressure gradient is given by $dP/dx = \Delta p g dD/dx$ where $P$ is pressure, $\Delta p$ is the density contrast between crust and mantle and $g$ is the acceleration due to gravity. For horizontal flow the rate of change in crustal thickness is given by

$$\frac{dD}{dt} = \frac{d}{dx} \int_0^D u dz.$$  

(4)

[12] Together with the standard Navier-Stokes equation for creeping flow in one dimension (see NS01), analytical solutions for the velocity may be obtained if $n$ is an integer. Solution of (2) and (4) is straightforward if $n = 1$; in this case, the time constant $\tau_H$ for decay of sinusoidal topography is given by

$$\tau_H = \frac{4}{\pi^{1/2}} \frac{\exp(\frac{Q}{RT_{bs}})}{AK^2\Delta p g}$$

(5)

where $k$ is the wavenumber ($= 2\pi/\lambda$). The expression for the time constant under basal heating $\tau_v$ is identical except that the numerical constant is $1/2$. In this latter case $\varepsilon$ depends on the basal heat flux rather than $H$ (NS01).

### 3. Method

[13] Because the heat flux on Mercury is unknown, the heat generated per unit volume in the undepleted mantle was assumed to simply be a factor $C$ times the terrestrial value at 4 Ga B.P., with $C$ an adjustable parameter and using the terrestrial radiogenic concentrations of Sun and McDonough [1989]. Two end-member situations were investigated. In one (internal heating), all the heat producing elements were assumed to reside within the crust with a uniform distribution. In the other (bottom heating), the heat producing elements were assumed to be distributed uniformly throughout the crust and mantle.

[14] In general, equation (4) is not analytically tractable. Instead, the numerical method described in NS01 was used to calculate the change in crustal thickness with time, calculating the value of $\delta$ using equation (3) for the internally heated case. An initially sinusoidal surface topography with a wavelength of 1000 km and a peak-to-peak amplitude of 2 km was assumed. The topography was assumed to be supported by crustal thickness variations. Lateral effective viscosity variations and a time-dependent heat flux were incorporated, and the model was started at 4 Ga b.p. As a proxy for the relaxation time, the time at which the amplitude decayed to 10% of its initial value was used. The numerical solutions were verified against equation (5) in the newtonian ($n = 1$) case.

[15] Calculations were carried out assuming two rheologies: a dry plagioclase rheology [Rybach and Dresen, 2000], assuming dislocation creep ($n = 3$) was the rate-limiting mechanism; and a dry diabase rheology [Mackwell et al., 1995]. The mean surface temperature was assumed to be 400 K [Soter and Ulrich, 1967]. Other values are given in Table 1.

### 4. Results

[16] Figure 1 shows the concentration factor $C$ which gives a relaxation timescale of 100 Ma as a function of crustal thickness for the two rheologies. Because lower crustal flow rates decrease with declining topographic slope and declining heat fluxes, topography which survives the first ~ 100 Ma decays little thereafter (NS01). As expected, greater crustal thicknesses require lower radiogenic heat production to produce the same relaxation time. For the same value of $C$ the timescale is longer for internal heating than for bottom heating, and longer for diabase than for plagioclase. For internal heating, a crustal thickness of 300 km requires values of $C$ of $\leq 0.55$ and 0.7, respectively, for plagioclase and diabase, equivalent to surface heat fluxes of 16 and 20 mW m$^{-2}$. In the case of bottom heating, a crustal thickness of 300 km requires $C$ to be $\leq 0.35$ and 0.45, respectively, for plagioclase and diabase, corresponding to heat fluxes of 10 and 12 mW m$^{-2}$.

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**Table 1. Mercury Parameters Used in Model**

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Unit</th>
<th>Value</th>
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</thead>
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<tr>
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</tr>
<tr>
<td>mantle thickness</td>
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<tr>
<td>$K$</td>
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</tr>
<tr>
<td>$\Delta p$</td>
<td>kg m$^{-3}$</td>
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</tr>
<tr>
<td>$Q$</td>
<td>kJ mol$^{-1}$</td>
<td>plagioclase</td>
</tr>
<tr>
<td>$A$</td>
<td>MPa$^{-n}$s$^{-1}$</td>
<td>5 $\times$ 10$^{12}$</td>
</tr>
<tr>
<td>$n$</td>
<td></td>
<td>3</td>
</tr>
</tbody>
</table>

Plag is plagioclase.

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**Figure 1.** Combinations of crustal thickness and radiogenic concentration factor $C$ resulting in a relaxation timescale of 0.1 Ga for both internal and bottom heating. Solid line is for dry plagioclase, dashed line for dry diabase; parameters given in Table 1.
5. Discussion and Conclusions

Figure 2 shows the depth to the 700 K isotherm at 4 Ga b.p. for the same combinations of Cr and crustal thickness which produce relaxation times of 100 Ma. This isotherm was probably at 40 km or shallower at 4 Ga b.p. according to the fault scarp evidence (see above). Figure 2 therefore shows that, unless the crust is as strong as dry diabase and is mainly internally heated, the maximum crustal thickness must be \( \leq 200 \) km. Furthermore, to satisfy the faulting observations the value of \( C \) must be \( \geq 0.8 \). In short, the fault observations require relatively high heat fluxes; for topography to persist with such heat fluxes, the crust must be relatively thin.

At sufficient depths, plagioclase (if present in the crust) will react to clinopyroxene plus quartz. The resulting increase in density may be sufficient to render the crust negatively buoyant with respect to the underlying mantle and possibly provide an independent upper bound on likely crustal thicknesses [Dupeyrat and Sotin, 1995]. The depth on Mercury at which the reaction is complete varies with temperature but is about 160 km at 1300 K [Wood, 1987].

Lateral crustal thickness contrasts give rise to so-called buoyancy forces [Molnar and Lyon-Caen, 1988] which vary as the square of the topographic contrast. These buoyancy forces will produce stresses over the elastic portion of the lithosphere. Assuming for simplicity that the elastic thickness is also represented by the depth to the 700 K isotherm (see Figure 2), the resulting stresses may be calculated. For 1 km of topographic contrast, the stresses are in the range 40–60 MPa for the maximum crustal thicknesses shown in Figure 2. These stresses are perhaps a factor of 4–6 larger than typical stress drops observed in earthquakes on Earth [Scholz, 1982]. If the crust on Mercury really is 100–200 km thick, faults on Mercury must be stronger than terrestrial faults. It has been argued elsewhere that faults on Venus are strong because of the absence of water [Foster and Nimmo, 1996]; it is likely that faults on Mercury are strong for the same reason.

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References


Nimmo, F., and D. J. Stevenson, Estimates of Martian crustal thickness


F. Nimmo, Department of Geological Sciences, University College London, Gower St, London, WC1E 6BT, UK. (nimmo@gps.caltech.edu)