Constraining the crustal thickness on Mercury from viscous topographic relaxation

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[1] Mercury exhibits long-wavelength topography which has probably survived for ~4 Ga. Assuming Airy compensation, the survival of the topography indicates that only certain combinations of crustal thickness and thermal structure are allowable. A dry diabase rheology allows a thicker crust than a dry plagioclase rheology. The existence of ancient faults places some constraints on the thermal structure. Unless the crust of Mercury is both as strong as dry diabase and is heated mainly from within, the crustal thickness must be ≤200 km. The faulting evidence implies that the concentration of radiogenic elements in the crust plus mantle of Mercury is at least 80% of the terrestrial value. Combined with previous studies of the long wavelength gravity and topography, the crustal thickness of Mercury is probably 100–200 km. Faults on Mercury are probably several times stronger than terrestrial faults. *INDEX TERMS:* 5430 Planetology: Solid Surface Planets: Interiors (8147); 5418 Planetology: Solid Surface Planets: Heat flow; 6235 Planetology: Solar System Objects: Mercury; 8159 Tectonophysics: Rheology—crust and lithosphere

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 [*Vilas*, 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 (T_e) 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 material [*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

$$T(z) = T_s + \frac{H}{2K}\left(D^2 - z^2\right) \qquad (1)$$

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Table 1. Mercury Parameters Used in Model

Quantity	Unit	Value	
g	m s^{-2}	3.76	
radius	km	2440	
mantle thickness	km	610	
K	Wm^{-1} K^{-1}	3	
$\Delta\rho$	kg m^{-3}	500	
		plag.	diabase
Q	kJ mol^{-1}	648	500
A	MPa^{-n} s^{-1}	5×10^{12}	4.9×10^{3}
n		3	4.5

Plag is plagioclase.

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

$$\dot{\varepsilon} = \frac{\partial u}{\partial z} = A|\tau|^{n-1}\tau e^{-(Q/RT)} \sim Be^{-(z/\delta)^2} \quad (2)$$

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

$$\delta = T_b\left(\frac{2KR}{QH}\right)^{1/2} \quad (3)$$

where T_b 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 $dP/dx = \Delta\rho g\ dD/dx$ where P is pressure, $\Delta\rho$ 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

$$\frac{dD}{dt} = \frac{d}{dx}\int_0^D u\, dz. \quad (4)$$

[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 τ_H for decay of sinusoidal topography is given by

$$\tau_H = \frac{4}{\pi^{1/2}}\frac{\exp(Q/RT_b)}{Ak^2\delta^3\Delta\rho g} \quad (5)$$

where k is the wavenumber ($= 2\pi/\lambda$). The expression for the time constant under basal heating τ_F is identical except that the numerical constant is 1/2. In this latter case δ 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 δ 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 [*Rybacki 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 $\sim$ 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}.

[17] 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}.

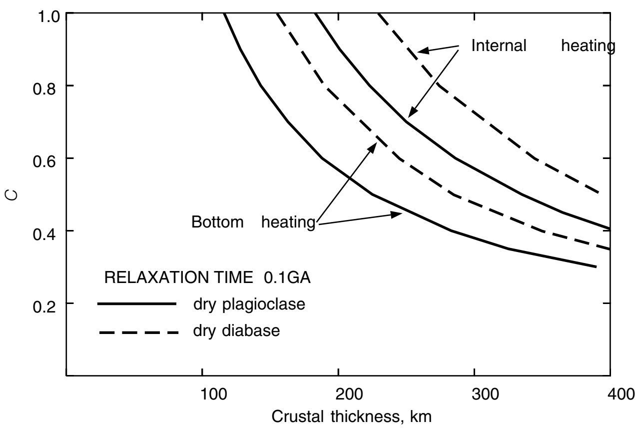

Figure 1. Combinations of crustal thickness and radiogenic concentration factor C resulting in a relaxation timescale of 0.1 Ga for both internal and bottom heating. Solid line is for dry plagioclase, dashed line for dry diabase; parameters given in Table 1.

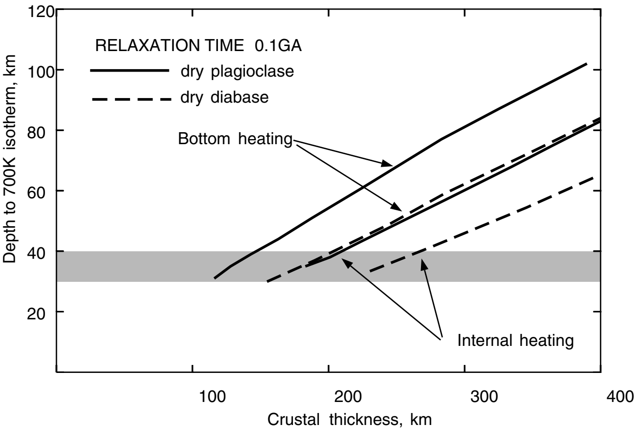

Figure 2. Depth to 700 K isotherm for same combinations of C and crustal thickness shown in Figure 1. Shaded box denotes approximate depth to 700 K isotherm determined from fault observations (see text).

[18] Figure 2 shows the depth to the 700 K isotherm at 4 Ga B.P. for the same combinations of C 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.

5. Discussion and Conclusions

[19] The main conclusion of this work is that, if the faulting observations provide a reasonable guide to heat fluxes on Mercury at 4 Ga b.p., topography is most likely to have survived if the mean crustal thickness were less than 200 km. If the *Anderson et al.* [1996] crustal thickness estimate of 100–300 km is correct, the crustal thickness on Mercury is probably 100–200 km. The faulting observations also imply that the overall concentration of radiogenic elements in the crust plus mantle of Mercury is at least 80% of terrestrial values.

[20] The greatest uncertainty in the above estimates is from the unknown composition of Mercury's crust. Dry diabase is nearly as strong as mantle materials [*Mackwell et al.*, 1995] and therefore provides a reasonable upper bound on likely crustal thicknesses. If present, weaker materials, such as dry plagioclase, provide tighter constraints on the maximum crustal thickness.

[21] Despite the uncertainty in rheology, the maximum crustal thickness of 200 km is probably an upper bound for at least four reasons. Firstly, it ignores the effect of secular cooling, which would increase heat fluxes and reduce allowable crustal thicknesses. Secondly, because high concentrations of radiogenic elements are generally produced by small melt fractions, it is hard to produce a crust which is both thick and has radiogenic concentrations high enough to approach the internal heating end-member case. Thirdly, the seismogenic layer is unlikely to be determined by an isotherm colder than 700 K; hotter temperatures, such as those appropriate to terrestrial mantle material, would reduce the acceptable crustal thickness (see Figure 2). And finally, for values of $C \sim 1$, the base of the crust would melt if crustal thicknesses were as large as indicated in Figure 1. On the other hand, although the internally heated and bottom heated situations are likely end members, there are possible situations (e.g. a highly enriched near-surface layer) which could result in sustainable topography for greater crustal thicknesses.

[22] 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].

[23] 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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