Thickness of the Martian crust: Improved constraints from geoid-to-topography ratios

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[1] The average crustal thickness of the southern highlands of Mars was investigated by calculating geoid-to-topography ratios (GTRs) and interpreting these in terms of an Airy compensation model appropriate for a spherical planet. We show that (1) if GTRs were interpreted in terms of a Cartesian model, the recovered crustal thickness would be underestimated by a few tens of kilometers, and (2) the global geoid and topography signals associated with the loading and flexure of the Tharsis province must be removed before undertaking such a spatial analysis. Assuming a conservative range of crustal densities (2700–3100 kg m^{-3}), we constrain the average thickness of the Martian crust to lie between 33 and 81 km (or 57 ± 24 km). When combined with complementary estimates based on crustal thickness modeling, gravity/topography admittance modeling, viscous relaxation considerations, and geochemical mass balance modeling, we find that a crustal thickness between 38 and 62 km (or 50 ± 12 km) is consistent with all studies. Isotopic investigations based on Hf-W and Sm-Nd systematics suggest that Mars underwent a major silicate differentiation event early in its evolution (within the first ~30 Ma) that gave rise to an "enriched" crust that has since remained isotopically isolated from the "depleted" mantle. In comparing estimates of the thickness of this primordial crust with those obtained in this study, we find that at least one third of the Martian crust has an origin dating from the time of accretion and primary differentiation. Subsequent partial melting of the depleted mantle would have given rise to the remaining portion of the crust. While we predict that a large portion of the crust should be composed of ancient "enriched" materials, a representative sample of this primordial crust does not currently exist among the known Martian meteorites. *INDEX TERMS:* 6225 Planetology: Solar System Objects: Mars; 5417 Planetology: Solid Surface Planets: Gravitational fields (1227); 5430 Planetology: Solid Surface Planets: Interiors (8147); 1227 Geodesy and Gravity: Planetary geodesy and gravity (5420, 5714, 6019); *KEYWORDS:* Mars, crustal thickness, crustal evolution, geoid, topography

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1. Introduction

[2] All of the terrestrial planets are known to possess a relatively low-density crust that is chemically distinct from primitive solar system objects (i.e., the chondritic meteorites). The formation of this crust is the result of early differentiation processes, as well as the cumulative effects of the planet's magmatic evolution over the past 4.5 billion years. For example, the bulk of the lunar crust is believed to have formed by the flotation of plagioclase in a crystallizing magma ocean roughly 4.5 billion years ago, with subsequent partial melting of the mantle only giving rise to volumetrically minor additions of basaltic magmas. In contrast to the Moon, the Earth does not currently posses a "primary" crust dating from the time of accretion. Instead, oceanic crust is currently being generated at oceanic spreading centers, and processes related to subduction have slowly given rise to a more complex continental crust.

[3] A knowledge of the crustal thickness of a planetary body can be used to constrain (among others things) the magmatic processes responsible for its formation, and the bulk composition and origin of the planet. For the Earth and Moon, these numbers are fairly well known and are thus commonly used in such studies. For Mars, however, published estimates of its mean crustal thickness have ranged from approximately 1 km to more than 250 km. As the crustal thickness is a key parameter used in the interpretation of the moment of inertia of Mars, a more precise estimate would help refine the radius of the Martian core,

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Table 1. Preferred Limits of Martian Crustal Thickness

Technique	Crustal Thickness, H_0, Limits, km	Reference
Assuming Hellas is Isostatically Compensated	>29	this study
Localized Admittances Using Technique of *Simons et al.* [1997]		
Western Hellas Rim	50 (+18, −24)	*McGovern et al.* [2002]
Hellas	50 (+12, −12)	*McGovern et al.* [2002]
Noachis Terra	50 (+12, −42)	*McGovern et al.* [2002]
Crustal Thickness Inversions Assuming a Minimum Crustal Thickness of 3 km	>32	this study
Viscous Relaxation of Dichotomy Boundary Combined with Crustal Thickness Inversions	<100	*Zuber et al.* [2000]; *Zuber* [2001]
Viscous Relaxation of Dichotomy Boundary and Hellas Basin	<115	*Nimmo and Stevenson* [2001]
Thorium Mass balance	<93 km	this study
Geoid-to-Topography Ratios for the Southern Highlands	57 ± 24	this study

and thus constrain whether or not a perovskite-bearing zone exists in the deep mantle [*Sohl and Spohn*, 1997; *Bertka and Fei*, 1998a; *Kavner et al.*, 2001]. The existence of such a perovskite layer has been shown to affect the nature of convection in the Martian interior, and may be responsible for the long-lived nature of the Tharsis and Elysium "plumes" [*Harder and Christensen*, 1996; *Harder*, 1998, 2000]. In addition, a recent thermal evolution study by *Hauck and Phillips* [2002] has highlighted the fact that the thickness of the Martian crust can help discriminate between different estimates of the bulk concentration of heat-producing elements in this planet, as well as the rheology of the mantle which is heavily influenced by the presence of H_2O.

[4] One technique for constraining the crustal thickness of a planetary body is to model the relationship between its observed gravity and topography fields. A number of such studies have been performed, especially since the Mars Global Surveyor (MGS) mission dramatically improved our knowledge of both the gravity field [*Lemoine et al.*, 2001; *Yuan et al.*, 2001] and topography [*Smith et al.*, 1999, 2001] of the planet. However, as we review below, these estimates are usually only strictly valid for specific regions of Mars and often involve a number of assumptions that must be carefully considered before extrapolating to the rest of the planet. In this study, we have calculated geoid-to-topography ratios (GTRs) over the areally extensive ancient southern highlands of Mars and have interpreted these in terms of the spectrally weighted admittance model of *Wieczorek and Phillips* [1997]. This technique accounts for the spherical nature of Mars, and we show that significant error would result if one were instead to use a Cartesian formalism. Furthermore, we show that before performing such a "spatial" analysis in the highlands, that one must first remove the global gravity and topography signature that is associated with the loading and flexure of the Tharsis province [*Phillips et al.*, 2001]. Assuming that the highlands are compensated by an Airy mechanism, we obtain a mean Martian crustal thickness between 33 and 81 km (or 57 ± 24 km). When this number is combined with previous complementary results, we find that the mean thickness of the Martian crust should lie between 38 and 62 km (or 50 ± 12 km).

[5] In this paper, we first review the literature associated with constraining the thickness of the Martian crust. Following this, we discuss the methodology that we use to interpret GTRs on a sphere and validate our approach using synthetic data. Next, we present and interpret our crustal thickness determinations. Finally, we discuss the implications of these results in terms of the thermal evolution of Mars and the origin of its crust.

2. Review and Critique of Previous Crustal Thickness Determinations

[6] In the absence of definitive Martian seismic data [cf. *Anderson et al.*, 1977], a number of techniques have been used in order to indirectly constrain the mean thickness of the Martian crust (H_0). These studies include analyses of Martian gravity and topography data, the moment of inertia of Mars, the viscous relaxation of topography, and geochemical mass balance calculations based on the composition of the Martian meteorites and Mars Pathfinder soils. As each of these methods employs a variety of assumptions, we here review and critique the techniques that have been used to date. Those estimates that we ultimately feel are the most reliable are summarized in Table 1.

2.1. Global Gravity and Topography Admittance Studies

[7] The gravity and topography fields of Mars have traditionally been expanded in terms of spherical harmonics, as these are the natural basis functions for a sphere. If one assumes that the entire planet is compensated by the same mechanism, then the ratio between the spectral components of these two fields (the admittance) as a function of degree l can be used to estimate a number of geophysical quantities, including the mean crustal thickness, the effective elastic thickness, and crustal density. For instance, using Viking-era gravity and topography models, and assuming that the crust was isostatically compensated by an Airy mechanism (i.e., mountains possess low-density crustal roots), *Bills and Nerem* [1995] estimated the mean crustal thickness to lie between 50 and 200 km (for $4 < l < 40$). Using improved MGS data, *Yuan et al.* [2001] showed that the crustal thickness must be greater than 100 km. However, when the gravitational signature of the Tharsis volcanoes was removed in this latter study, H_0 was found to lie between 50 and 200 km (for $10 < l < 60$, see their Figure 4b).

[8] These studies implicitly require that each spherical-harmonic degree l is compensated by a single mechanism, and furthermore that the density of the crust is uniform. These conditions, however, are most likely not satisfied when applied to Mars as a whole: the ice caps clearly have a reduced density, and many volcanic and impact structures

are at least partially elastically supported by the lithosphere. While the *Yuan et al.* [2001] study attempted to remove the largest non-isostatic effects associated with young volcanic features, some non-isostatic signatures are likely to remain and it is difficult to quantify how this would affect these crustal thickness estimates.

2.2. Spatial Gravity and Topography Studies

[9] The topography of Mars is clearly being supported by several different mechanisms, and by analyzing the gravity and topography of individual regions it is possible to bypass some of the problems associated with global admittance studies. For instance, using Mariner 9-era gravity and topography data, *Phillips and Saunders* [1975] argued that the ancient southern highlands of Mars were largely compensated based on the near-zero free-air gravity anomalies found there. In contrast, younger regions, such as the Tharsis plateau, were found to be inconsistent with this hypothesis. By assuming Airy compensation for the southern highlands, and minimizing the difference between the observed and predicted gravity fields, they constrained the mean crustal thickness to be less than 100 km. However, as the gravity model that was employed was then only known to l = 8, and it is now known that loading and flexure associated with the Tharsis province dominates the gravity field up to at least degree six [e.g., *Zuber and Smith*, 1997; *Phillips et al.*, 2001] it is not clear if this result is entirely valid. Using improved Viking-era gravity and topography models, *Frey et al.* [1996] confirmed that large regions of the southern highlands were indeed isostatically compensated. While this study also estimated depths of compensation, the technique employed therein was never validated.

[10] *Sjogren and Wimberley* [1981] used Viking I line-of-sight gravity data to model the crustal structure beneath the Hellas impact basin. Using a combination of three spacecraft profiles over the northern portion of this crater, the observed data were best fit by a 135 ± 21 km thick crust that possessed an uplifted mantle, a structure similar to many lunar impact basins [e.g., *Bratt et al.*, 1985; *Neumann et al.*, 1996; *Wieczorek and Phillips*, 1999]. This estimate is unlikely to be accurate as a result of the poor resolution of the then available topography model (a rim-to-floor elevation difference of only 3 km was used for this basin, in comparison to the presently known value of ~8 km). The mean thickness of the crust in the region of the 370-km diameter crater Antoniadi was similarly estimated by *Sjogren and Ritke* [1982] using a single line-of-sight gravity profile in combination with then currently available high-resolution topographic profiles for this region. Assuming this crater to be isostatically compensated, the mean thickness of the crust in this region was estimated to lie between 104 and 126 km. However, the simplified model of the gravity inversion (only six mass-disks were used), the unqualified assumption of complete isostasy for this medium sized crater, and the limited number of topographic profiles limited the robustness of this result.

[11] Assuming that a region is isostatically compensated, the linear relationship between its geoid and topography (i.e., the geoid-to-topography ratio, or GTR) can be used to infer the mean thickness of the crust [*Ockendon and Turcotte*, 1977; *Haxby and Turcotte*, 1978]. Using such an approach and employing MGS-derived gravity and topography data, *Turcotte et al.* [2002] have recently inferred a mean crustal thickness of 90 ± 10 km from a single profile that spans the Hellas basin. However, while the use of GTRs is well established for the Earth, *Wieczorek and Phillips* [1997] have shown that considerable error can incur when this technique is applied to small planets with significant curvature such as the Moon. In section 3, we show that the Cartesian approach used by *Turcotte et al.* [2002] underestimates the true crustal thickness of Mars by tens of kilometers. We further show that it is necessary to remove the global gravity and topography signatures associated with the loading and flexure of the Tharsis province before performing such a spatial analysis.

[12] If the topography of Mars is everywhere isostatically compensated, and if the density of the crust and mantle are known, then a minimum estimate of H_0 can be obtained by requiring the minimum thickness of the crust to be equal to zero. Since the lowest elevations of Mars are found within the Hellas basin, the thickness of the crust in this region might be expected to be the thinnest as well (though see section 2.4). Using this approach, *Nimmo and Stevenson* [2001] reported a minimum mean crustal thickness of Mars of ~30 km. However, they used a rather restrictive range of crustal densities (2800–2900 kg m^{-3}) and a range of mantle densities that is probably too low (3300–3400 kg m^{-3}). They further appear to have used the average elevation of the Hellas floor in their calculations, whereas the absolute minimum might be better given the possible presence of partially uncompensated sediments in this basin [e.g., *Moore and Wilhelms*, 2001]. Given these concerns, we redo this calculation here. Using a 90th degree spherical harmonic shape model of Mars (and reducing J_2 by 94% in order to account for the rotational flattening of Mars [*Folkner et al.*, 1997; *Zuber and Smith*, 1997]), we obtain a minimum elevation in the Hellas basin of −6.81 km. As the density of the southern highland crust is not well constrained, we conservatively use the range 2700–3100 kg m^{-3}, whereas for the density of the upper mantle we use the range of 3400–3550 kg m^{-3} [e.g., *Sohl and Spohn*, 1997; *Bertka and Fei*, 1998a], which is based on the mantle compositional model of *Dreibus and Wänke* [1985]. Using these density limits we find that H_0 must be greater than 29 km, in agreement with the previous result. Using the more restrictive density limits of *Nimmo and Stevenson* [2001] would increase this minimum crustal thickness to 39 km.

2.3. Local Gravity and Topography Admittance Studies

[13] Whereas spatial studies of localized regions usually only result in a single degree-independent admittance, spectral studies of localized regions yield a wavelength-dependent admittance function that can often be inverted for more than one model parameter. Two general techniques have been used to calculate localized admittances on a planet. One technique employs a spherical harmonic model of both gravity and topography and localizes specific regions by multiplying these fields by a windowing function. The other method is to directly model the line-of-sight spacecraft accelerations.

[14] *McGovern et al.* [2002] has employed the former technique using the localization method of *Simons et al.*

[1997]. A benefit of this approach is that admittances are calculated and interpreted directly in terms of spherical harmonics. Nevertheless, because of the windowing methodology involved, there exists a trade-off between spatial and spectral resolution that limits the shortest and longest wavelengths that can be analyzed. Their analysis included volcanic structures, impact craters, and the southern highlands, and the obtained admittances were interpreted in terms of an elastic loading model that included both surface and subsurface loads (which were assumed to be in phase). This study obtained estimates of the elastic thickness for these features, and the crustal thickness was found to be best constrained where the computed elastic thicknesses were low (i.e., where the crust was close to an isostatic state).

[15] One such region is Noachis Terra in the southern highlands. Here, the best fit elastic thickness was found to be zero (with a maximum allowable value of 25 km), the mean crustal thickness was constrained to be 50 (+12, −42) km, and a best fit crustal density of 2700 kg m^{-3} was determined. For the western rim of the Hellas basin an elastic thickness of 5 (+14, −5) km was obtained, with a mean crustal thickness of 50 (+18, −24) km. And finally for the Hellas basin itself, the elastic thickness was constrained to be 5 (+7, −5) km with a crustal thickness of 50 (+12, −12) km. These results indicate that the ancient southern highlands are close to being in a state of isostatic equilibrium, and limit the mean crustal thickness of Mars to lie between 8 and 68 km (with a best fit of 50 km).

[16] Two admittance studies using line-of-sight data have been performed to date. *McKenzie et al.* [2002] calculated 1-dimensional Cartesian admittances along individual spacecraft profiles over the southern hemisphere, Tharsis, Valles Marineris and Elysium, and interpreted these in terms of a simple elastic-plate model with surface loading. Only for the southern hemisphere were they able to constrain the effective depth of compensation, in which they obtained a value of <10 km with an associated elastic thickness of 14.5 km. As they acknowledged, this depth of compensation is unlikely to represent the true crustal thickness for this region given the high elevations that are found there. Instead, it was suggested that this depth might instead represent the thickness of a low-density impact-brecciated surface layer. However, given the large size of this study region (all latitudes southward of 20°S), several compensation mechanisms could plausibly be operating there (for example, the uncompensated Argyre "mascon" basin, Airy compensated highlands, and flexural support of the south-polar ice cap). It is thus challenging to interpret their calculated effective depth of compensation and elastic thickness. Furthermore, as coherences were not calculated in this study, the extent to which a model with only surface loads is appropriate it is not entirely clear [e.g., *Forsyth*, 1985].

[17] Using a similar technique, *Nimmo* [2002] calculated 1-dimensional Cartesian line-of-sight admittances over the region of the dichotomy boundary. These results were interpreted in terms of an elastic-plate loading model that possessed two crustal layers, and where surface and crust-mantle interface loads possessed random phases. The mean crustal and elastic thicknesses obtained for this region were found to lie in the range of 1–111 km and 21–113 km, respectively. By fixing the density of the upper-crustal layer, it was possible to reduce the crustal thickness range to 1–75 km and the elastic thickness range to 37–89 km. Coherences were calculated in this study, but were never compared to the predictions of the employed loading model. Given that the calculated coherences are extremely low at both long and short wavelengths (~0.1 for $\lambda >$ 700 km and $\lambda <$ 350 km), with a maximum value of about 0.65 at wavelengths near 400 km, it is unclear whether the employed admittance model is entirely appropriate for this region. As a result of the unique geology associated with the dichotomy boundary, interpretations of admittance results for this region are also hindered by the likely presence of multiple compensation mechanisms that could be operating there (for example, Airy compensated highlands, uncompensated sediments in the northern lowlands, and a possible change from Airy to Pratt compensation across the dichotomy boundary).

2.4. Global Crustal Thickness Inversions

[18] One of the simplest interpretations of a planet's gravity field (with the exception of the J_2 rotational flattening) is that it is exclusively the result of both surface topography and relief along the crust-mantle interface. If one further assumes constant values for the density of the crust and mantle, as well as a mean crustal thickness, then it is straightforward to invert for the relief along the "Moho," thus providing a global crustal-thickness map [e.g., *Wieczorek and Phillips*, 1998]. As the average thickness of the crust is in general not known, the thickness of the crust is often "anchored" to a specific value at a given locality in these inversions. For the Moon, this has been done by using the seismically constrained crustal thickness beneath the Apollo 12 and 14 landing sites [cf. *Khan and Mosegaard*, 2002; *Lognonné et al.*, 2003]. In the absence of definitive seismic data for Mars [cf. *Anderson et al.*, 1977], an alternative approach is to arbitrarily set the minimum crustal thickness to a value near zero, thus providing a minimum estimate of the mean crustal thickness.

[19] Such an approach was originally employed by *Bills and Ferrari* [1978] using Viking-derived gravity and the best available topography data. The crustal thickness inversions in this study showed that the thinnest crust should be located beneath the Hellas impact basin. By constraining the minimum crustal thickness to be zero there, a minimum globally averaged crustal thickness of 23–32 km was obtained. The results from this study however suffered from the low resolution of the then available gravity and topography models that were known only to a maximum degree of 16 and 10, respectively.

[20] Using MGS-derived gravity and topography data, *Zuber et al.* [2000] constructed improved crustal thickness maps valid to approximately spherical-harmonic degree 60. In contrast to *Bills and Ferrari* [1978], they found that the minimum crustal thickness occurred not beneath the Hellas basin, but beneath Isidis. Arbitrarily setting the minimum crustal thickness there to 3 km (and using crustal and mantle densities of 2900 and 3500 kg m^{-3}) a mean crustal thickness of 50 km was obtained. As they noted, this value should be used as a minimum estimate as the true crustal

thickness beneath Isidis could in fact be much larger. It was later found, however, that this mean crustal thickness value was mistakenly referenced to the mean equatorial radius, as opposed to the mean planetary radius. Correcting for this, this value should be revised downward to 44 km. While the crustal and mantle densities used by *Zuber et al.* [2000] probably represent the best globally averaged estimates for these values, uncertainties associated with these parameters will inevitably affect the estimated mean crustal thickness. We have thus re-performed these calculations using a wider range of allowable crustal (2700–3100 kg m^{-3}) and mantle densities (3400–3550 kg m^{-3}) and find that the mean crustal thickness of Mars should be greater than 32 km.

2.5. Studies That Utilize Moment-of-Inertia Constraints

[21] The moment of inertia of a planet provides constraints on its radial density profile and can thus provide limits on both the thickness of the crust and the radius of its core. As the inferred core radius and crustal thickness depend on their assumed densities, as well as the assumed mantle density profile, the moment of inertia by itself provides little quantitative information. Nevertheless, by applying additional constraints, it is possible to limit the range of solutions. One such approach is to calculate a mantle density profile based on geochemical arguments and to then use the moment of inertia to constrain the properties of the crust and core.

[22] By employing the mantle and core composition model of *Dreibus and Wänke* [1985] (see *Longhi et al.* [1992] for a review), *Sohl and Spohn* [1997] theoretically calculated the Martian mantle density profile and generated two end-member models of interior structure. The first model assumed a bulk chondritic Fe/Si ratio for the planet (which is an explicit assumption of the *Dreibus and Wänke* [1985] model), and obtained a crustal thickness of 252 km. While the calculated moment of inertia of this model was within the range of the then available estimates, it is inconsistent with the present estimate based on Mars Pathfinder data [*Folkner et al.*, 1997]. The other model did not fix the bulk Fe/Si ratio, and was constrained to match the most reliable of the pre-Mars Pathfinder moment-of-inertia estimates (which lies within the error limits of the presently known value). A mean crustal thickness of 110 km was determined for this model, but when a range of crustal densities were considered (2600–3200 kg m^{-3}), permissible crustal thickness was found to vary between 50 and 200 km. If the uncertainties associated with the present value of the moment of inertia were to be taken into account, or if the core composition was allowed to deviate from that assumed by *Dreibus and Wänke* [1985], then the range of crustal thickness estimates would be further increased.

[23] *Bertka and Fei* [1998a] performed a similar set of calculations using the mantle composition model of *Dreibus and Wänke* [1985]. In their study, the experimentally determined modal mineralogy of the mantle [*Bertka and Fei*, 1997] was used to compute a mantle density profile, and the core radius and crustal thickness were then determined by employing mass and moment-of-inertia constraints. In agreement with *Sohl and Spohn* [1997], they found that if the bulk composition of Mars was constrained to have a chondritic Fe/Si ratio, a large crustal thickness was required (between 180 and 320 km). As was noted, however, this model is inconsistent with the presently known moment of inertia. Using the *Dreibus and Wänke* [1985] core composition (but not constraining the bulk planetary composition) and assuming crustal densities between 2700 and 3000 kg m^{-3}, the thickness of the crust was required to be greater than 42 km. In a different study, *Bertka and Fei* [1998b] used the Mars Pathfinder moment-of-inertia constraint, and considered a much wider range of possible core compositions. While allowable crustal thicknesses were found to lie between 35 and 80 km, this study did not consider the uncertainties associated with the moment of inertia, and further assumed a single crustal density of 3000 kg m^{-3}.

[24] In a study by *Kavner et al.* [2001], improved experimental constraints were placed on the density and phase diagram of Fe-FeS alloys, showing that the solid phase FeS(V) should be stable at Martian core pressures. Using these new density measurements (along with the *Bertka and Fei* [1998a] mantle density profile) and assuming a crustal density of 2800 kg m^{-3}, the allowable range of crustal thickness, core radius, and core composition were investigated. Assuming that the core is more iron rich than FeS, the crustal thickness was constrained to be less than 125 km. However, as this study inadvertently used the polar moment of inertia C (0.3662 ± 0.0017) for the average moment of inertia I (0.3649 ± 0.0017), their maximum crustal thickness should be revised upward by about 25 km. Considering a wider range of crustal densities would further increase this upper bound.

[25] One potentially important concern that is common to the studies discussed above (as well as to most others in the literature) is that the equations of state used for the Martian core are only strictly applicable for solid phases (i.e., FeS(IV), FeS(V), and γ-iron), whereas recent evidence exists that the core is at least partially molten [*Yoder et al.*, 2003]. This drastic simplification has been partly used out of necessity due to the near-absence of experimental data for iron-sulfur liquids at high pressure. As the density of pure liquid and solid iron are expected to be close at high pressures (within a few percent [e.g., *Jeanloz*, 1979; *Longhi et al.*, 1992]), this was generally thought to be a reasonable approximation. However, recent experiments performed on iron-sulfur liquids by *Sanloup et al.* [2000, 2002] indicate that the compressibility of these liquids is heavily dependent on its sulfur content. For example, the isothermal bulk modulus K_T was found to decrease from about 85 GPa for pure iron to about 11 GPa for an alloy with 27 wt.% sulfur. As discussed by *Sanloup et al.* [2000, 2002], this could have important consequences on moment-of-inertia calculations when even small amounts of sulfur are present in the core.

2.6. Viscous Relaxation of Surface Topography Studies

[26] Even if the Martian crust was completely isostatically compensated, lateral variations in crustal thickness would still give rise to lateral pressure gradients within the crust and upper mantle. As rocks behave viscously over long periods of time, these pressure gradients would cause the crust to flow, with the velocity of this flow being roughly inversely

proportional to the material's viscosity. Since temperature increases with depth below the surface, most crustal flow will occur in the lower portions of the relatively hotter crust. Moreover, as the thickness of the crust increases, the temperature of the lower crust will increase, reducing the time required for significant viscous relaxation to occur. Two extreme cases illustrate this effect. If the crust were extremely thick, then high rates of viscous flow due to high temperatures would quickly remove any pre-existing crustal thickness variations. In contrast, if the crust were extremely thin, the colder temperatures encountered there would give rise to a high viscosity, thus allowing crustal thickness variations to survive over the age of the planet. One can thus ask the question: Can the *existence* of crustal thickness variations and surface topography be used to constrain the maximum thickness of the crust?

[27] This question was addressed in two studies. As discussed above, *Zuber et al.* [2000] used recently obtained gravity and topography data for Mars to construct a global crustal thickness model with a minimum average thickness of about 50 km. As studies based on moment-of-inertia and geochemical considerations suggested that the average crustal thickness might in fact be much larger (section 2.5), it was possible that this model significantly underestimated the true thickness of the Martian crust. Nevertheless, by using the viscous relaxation model of *Zhong and Zuber* [2000], they argued that the pole-to-pole crustal thickness gradient could not have been maintained over geologic time if the average thickness of the crust was greater than $\sim$100 km.

[28] *Nimmo and Stevenson* [2001] carried out a more detailed set of relaxation calculations for the north-south dichotomy boundary and the Hellas basin. As in the study of *Zuber et al.* [2000], the models depend on many poorly known parameters, including the viscosity of the crust, the density contrast between the crust and mantle, the crustal temperature profile, and the initial assumed topography. Among these, uncertainties associated with the assumed temperature profile were found to have the largest effect on their calculations. Nevertheless, for the model that possessed the coldest temperature profile, it was found that the average crustal thickness could not exceed 115 km. For more reasonable temperature profiles, a maximum crustal thickness of about 100 km was obtained, in agreement with the study of *Zuber et al.* [2000].

2.7. Potassium, Thorium, and Uranium Mass Balance Models

[29] One manner in which the thickness of a planet's crust can be determined is by geochemical mass balance considerations. In these types of models a planet is divided into a discreet number of uniform geochemical reservoirs from which the bulk planetary composition can be calculated. One of the simplest cases is to assume that the mass of the silicate portion of the planet is known, and that it further consists solely of crustal and mantle materials. If the composition of these two silicate reservoirs was known, then with a knowledge of the bulk silicate composition (usually assumed to be chondritic) the mass of the crust would be uniquely determined.

[30] One such mass balance calculation has been carried out by *McLennan* [2001] for the long-lived heat-producing elements K, Th, and U. In particular, a bulk potassium concentration of the Martian crust was determined on the basis of the composition of soils at the Mars Pathfinder site [*Wänke et al.*, 2001] and data from the Phobos-2 gamma-ray spectrometer [*Surkov et al.*, 1989, 1994; *Trombka et al.*, 1992]. The bulk silicate abundances of these elements, including their K/U and K/Th ratios, were taken from the geochemical model of *Dreibus and Wänke* [1985], and the crustal thickness was then solved for as a function of the percentage of radiogenic heat-producing elements present in the crust. In the extreme case where all heat-producing elements were sequestered in the crust, a maximum allowable crustal thickness of 65 km was found.

[31] As was acknowledged in the *McLennan* [2001] study, the uncertainties associated with this calculation are not easy to quantify. While his estimate of the bulk crustal concentration of potassium based on the Pathfinder soils agreed well with the Phobos-2 gamma-ray spectroscopy data, the abundance of potassium appears to be lower at the Viking landing sites (where only an upper bound was obtained [*Clark et al.*, 1982]). If the Viking measurements were representative of the Martian crust, then the above model would underestimate the true crustal thickness. In contrast, it was noted that the K/U and K/Th ratios as determined by the Phobos-2 mission (and corroborated by recent Mars Odyssey data [*Taylor et al.*, 2003]) appear to be smaller than those based on the Martian meteorites. Taking this into account in the above model would conversely give rise to higher Th and U abundances in the crust, and would thus give rise to a smaller calculated crustal thickness. In any case, if any amount of radioactive elements were present in the Martian mantle, then the modeled crustal thickness would be smaller yet.

[32] We further quantify this model by taking into account a larger range of crustal densities and core radii than were employed in the *McLennan* [2001] study. In particular, we have conservatively used a range of crustal densities between 2700 and 3100 kg m^{-3}, and core radii between 1150 and 1850 km [*Harder*, 1998]. An average mantle density of 3700 kg m^{-3} was used (based on the density profile of *Sohl and Spohn* [1997]), and the chondritic bulk silicate abundance of thorium was taken from the model of *Dreibus and Wänke* [1985] (0.056 ppm). Using the average crustal thorium abundance of *McLennan* [2001] (0.9 ppm), and assuming that all thorium is sequestered in the crust, we find the maximum allowable crustal thickness to be 93 km.

[33] Recent thermal evolution models performed by *Kiefer* [2003] indicate that the above assumption of zero heat production in the present Martian mantle is inconsistent with the presence of recent volcanic activity on Mars. In particular, he found that if the current heat production of the mantle was less than $\sim$40% of the bulk value of *Dreibus and Wänke* [1985], partial melting within a Tharsis plume would not occur. Geologic estimates of the recent magmatic production rate further constrained the heat production of the mantle to be less than $\sim$60% of the bulk *Dreibus and Wänke* [1985] value. Using these two limiting values on the abundance of thorium in the mantle, we find that the thickness of the Martian crust should lie between 29 and 57 km. While these limits are likely to be more realistic than the upper bound of 93 km reported above, the uncertainties

associated with the thermal model calculations are difficult to quantify.

2.8. Nd Isotopic and Mass Balance Studies

[34] One geochemical estimate of the thickness of the Martian crust has been made on the basis of a consideration of the neodymium isotopic compositions of the Martian meteorites. The basic premise of this modeling rests on the work of *Jones* [1989] and *Longhi* [1991], who argued that the isotopic composition of the shergottites (Martian basaltic meteorites) could be explained by mixing between two end-member compositions which they interpreted as "depleted" mantle and "enriched" crust. Indirect evidence for the existence of a primordial enriched crust comes from positive $\varepsilon^{142}{}_{\mathrm{Nd}}$ anomalies of mantle-derived partial melts which imply a major silicate differentiation event approximately 30 Ma into Mars history [*Harper et al.*, 1995; *Borg et al.*, 1997; *Lee and Halliday*, 1997; *Blichert-Toft et al.*, 1999]. (Other more indirect indications of early Martian differentiation are summarized by *Halliday et al.* [2001].) While the complement to this depleted mantle reservoir has not yet been sampled [e.g., *Lee and Halliday*, 1997; *Münker et al.*, 2003], it most plausibly resides in the crust which has since remained isotopically isolated from the mantle. This supposition is supported by elemental abundances derived from the Mars Pathfinder XPXS instrument [*Wänke et al.*, 2001], and the Phobos-2 [*Surkov et al.*, 1994] and Mars Odyssey gamma-ray spectrometers [*Taylor et al.*, 2003], which show that the bulk Martian crust is more enriched in incompatible elements than the Martian meteorites [see also *McLennan*, 2001].

[35] (Comments on Sm-Nd isotopes: Partial melting of a chondritic source will preferentially fractionate Sm into the solid phase. As the radioactive isotopes ^{147}Sm and ^{146}Sm decay into ^{143}Nd and ^{142}Nd with half-lives of 109 Ga and 103 Ma, respectively, the Nd isotopic evolution of the melt and restite will have separate time evolutions. For the "depleted" restite (or mantle), the $\varepsilon^{142}{}_{\mathrm{Nd}}$ and $\varepsilon^{143}{}_{\mathrm{Nd}}$ will be positive and increase with time. In contrast, for the "enriched" melt (or crust), these values will be negative and decrease with time.)

[36] Assuming that the depleted mantle and enriched crust were the only two isotopic reservoirs on Mars, *Norman* [1999] used their $\varepsilon^{143}_{\mathrm{Nd}}$ values to determine that 51.5% of the neodymium budget of Mars resides within the crust, with the remaining portion being sequestered in the mantle. Using this result, the thickness of the enriched crust was constrained by using a mass balance approach in the following manner. First, the isotopic composition of Shergotty was modeled as a binary mixture between an enriched crustal component, and a mantle-derived parental magma having a composition similar to EET79001. Second, by constraining the $\varepsilon^{143}{}_{\mathrm{Nd}}$ and Nd concentration of this mixture to be equal to that of Shergotty, the Nd concentration of the enriched crustal component was determined. Next, from knowing the total amount of Nd in the crust, its Nd concentration was calculated as a function of crustal thickness (this is simply the mass of neodymium divided by the mass of the crust). Finally, from the modeled crustal Nd abundance, limits on the crustal thickness were determined. Results from this model imply that the enriched crust of Mars is less than 45 km thick, with best-fit values lying between 20 and 30 km.

[37] As was emphasized by *Norman* [1999], the thickness of this "enriched crust" does not necessarily have to equal that of the "geophysical crust," the latter of which is based primarily on density considerations. The enriched crust in this model is essentially only a geochemical complement of the "depleted mantle." As such, these results depend on an accurate knowledge of the mantle and crustal neodymium abundances and isotopic compositions, as well as the likely assumption that the "enriched" component in fact resides within the crust. In reality, the diverse composition of the Martian meteorites suggests that the crust of Mars might be more adequately modeled as a mixture of both geochemically enriched and depleted components, the latter representing partial melts of a depleted mantle source. In this scenario, the above-calculated crustal thickness would represent only the thickness of the primordial crust, whereas subsequent melting of the depleted mantle source would have given rise to additional secondary crustal materials having a depleted isotopic signature [e.g., *Norman*, 2002].

[38] While it is encouraging that the enriched crustal-thickness estimates of *Norman* [1999] are similar to the geophysical estimates presented in Table 1, they should not be considered as one and the same because of the different definitions of "geophysical" and "geochemical" crust. Indeed, the geochemical primordial "enriched" crust should be thinner than the total crust (which will contain later magmatic additions from a depleted mantle). Nevertheless, as we discuss in section 5, these types of geochemical models give insight into the chemical composition and origin of the crust that could not otherwise be obtained from a solely geophysical investigation.

3. Crustal Thickness From Geoid-to-Topography Ratios: Methodology and Validation

[39] *Ockendon and Turcotte* [1977] and *Haxby and Turcotte* [1978] have shown that if a region is isostatically compensated, the geoid anomaly can be approximated by the dipole-moment of the vertical mass distribution. For the case of Airy isostasy, the geoid-to-topography ratio, or GTR, is to first order linearly related to the crustal thickness by the relationship

$$\mathrm{GTR} = \frac{N}{h} = \frac{2\pi\rho_c R^2}{M} H_0 \tag{1}$$

where N is the geoid height, h is the surface elevation, ρ_c is the crustal density, H_0 is the crustal thickness at zero elevation, and M and R are the mass and radius of the planet. (While M/R^2 can be replaced by the surface gravity, we write it in this form for comparison with the spherical equations below). This equation was derived under the assumption of Cartesian geometry and is only strictly valid in the long-wavelength limit.

[40] An improved technique for relating the measured GTR on a planet to an arbitrary model of crustal structure and compensation has been developed by *Wieczorek and Phillips* [1997]. This technique takes into account the spherical nature of planetary bodies, the wavelength-dependence of the assumed compensation model, and does not

require an assumption of isostasy. In particular, they showed that the geoid-to-topography ratio at a given point on the surface is to first order given by

$$\mathrm{GTR} = R \sum_{l} W_l Z_l \tag{2}$$

where Z_l is an arbitrary admittance function dependent on spherical harmonic degree l, and W_l is a degree-dependent weighting function given by

$$W_l = \frac{V_l^2}{\sum_l V_l^2}. \tag{3}$$

where V_l^2 is the topographic power at spherical harmonic degree l. Expressing the surface topography in spherical harmonics as

$$h(\theta, \phi) = \sum_{i,l,m} h_{ilm} Y_{ilm}(\theta, \phi) \tag{4}$$

the topographic power at degree l is given by

$$V_l^2 = \sum_{i,m} h_{ilm}^2. \tag{5}$$

For the case of Airy compensation, the admittance function is given by [e.g., *Lambeck*, 1988]

$$Z_l = \frac{C_{ilm}}{h_{ilm}} = \frac{4\pi\rho_c R^2}{M(2l+1)} \left[1 - \left(\frac{R-H}{R}\right)^l\right] \tag{6}$$

where C_{ilm} and h_{ilm} represent the spherical harmonic coefficients of the gravitational potential and surface topography, respectively. With a knowledge of the planet's topographic power spectrum, as well as an assumed crustal density, a theoretical GTR can be calculated as a function of crustal thickness using the above equations.

[41] As is evident from equations (2) and (3), the spatial GTR is simply a spectrally weighted sum of degree-dependent admittances, where the weighting function is proportional to the power of the topography. As it has been observed that planetary topographic power spectra are "red" (i.e., the topography is characterized by larger amplitudes at long-wavelengths), the GTR will be most heavily biased by the admittance function at long wavelengths (i.e., low values of l). For example, using the spherical harmonic representation of Martian surface topography [*Smith et al.*, 2001], and excluding those spherical harmonic degrees less than degree 10 (see section 4), we find that 50% and 90% of the GTR results from admittances with degrees less than 20 and 34, respectively.

[42] The technique of using spectrally weighted admittances to invert for crustal thickness was originally applied to the Moon by *Wieczorek and Phillips* [1997]. They found that if one was to apply the traditional Cartesian dipole-moment method to this body, that the inverted crustal thickness would always be underestimated, in some cases by more than 20 km. As the radius of Mars is about twice that of the Moon, the errors associated with assuming a Cartesian geometry for this body might be expected to be smaller. In Figure 1 we plot the predicted relationship between the GTR and crustal thickness for the two models, and it is seen that the Cartesian method still significantly underestimates the true crustal thickness of Mars. In particular, for a GTR of $\sim$15 m km^{-1} (which is representative of Mars), the Cartesian method underestimates the true value by about 15 km.

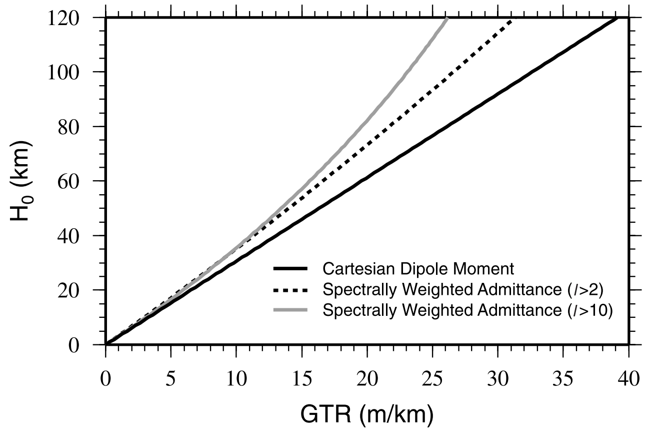

Figure 1. Predicted crustal thickness as a function of the geoid-to-topography ratio. Solid line represents the Cartesian model of *Ockendon and Turcotte* [1977] and *Haxby and Turcotte* [1978], whereas the dashed and gray lines represent the spectrally weighed admittance model of *Wieczorek and Phillips* [1997]. For the latter model, results are shown for two cases where different portions of the low-order topography harmonics were ignored. The crustal density was assumed to be 2900 kg m^{-3} and the maximum degree used in the topographic spherical harmonic expansion was 70.

[43] Before using the above method to infer limits on the crustal thickness of Mars, we first demonstrate its accuracy by inverting synthetic data. The rationale for doing this is twofold. First, the technique described above assumes that the geoid and topography are linearly related by a degree-dependent admittance function. While this is a good approximation when the relief along the surface and crust-mantle interface is small, this is no longer necessarily true when the relief is large [e.g., *Parker*, 1972; *Wieczorek and Phillips*, 1998]. Second, in this study, the GTR for a given location will be determined by regressing geoid and topography data within a circle of a given size, and we would like to determine how large such a circle must be in order to obtain reliable crustal thickness estimates.

[44] In testing this method, we have created a synthetic map of relief along the crust-mantle interface by assuming that the southern highlands of Mars are fully compensated with a zero-elevation crustal thickness of 50 km and a crustal and mantle density of 2900 and 3500 kg m^{-3}, respectively. The corresponding geoid anomaly due to the surface and crust-mantle interface was fully computed using the methodology of *Wieczorek and Phillips* [1998]. We note that while we have included the degree-1 topography in computing the synthetic geoid, that this term was later removed before computing the GTR as there is no corresponding degree-1 geoid signal in center-of-mass coordinates. For each point in the southern highlands (sampled on a 2° × 2° equal-area grid), geoid and topog-

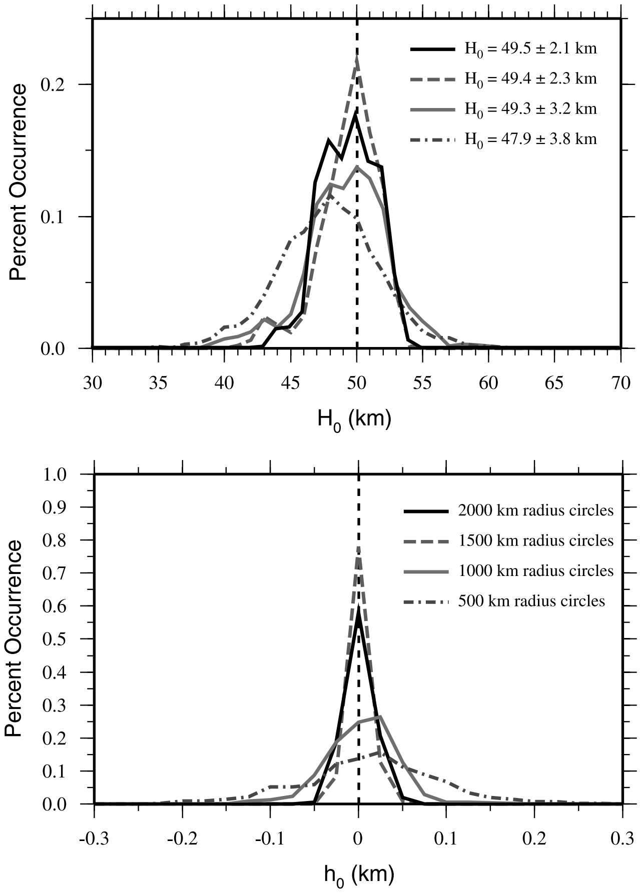

Figure 2. Recovered values of the zero-elevation crustal thickness, H_0 and the x-intercept, h_0, when circle radii of 500, 1000, 1500 and 2000 km were used to regress the geoid and topography data. The true crustal thickness for this synthetic test was 50 km.

raphy data lying within a circle of radius r were regressed according to the equation

$$N = \mathrm{GTR}(h - h_0) \tag{7}$$

in order to determine the GTR (i.e., the slope) and the x-intercept, h_0. If more than 25% of the data points within a given circle were absent (this occurs near the boundaries of the study region), or if the regression possessed more than a 1% probability of being uncorrelated, the region was ignored. For this test, GTRs were computed for the same regions that were used with the real data as is discussed further in the following section.

[45] From the obtained GTR for each region, the average crustal thickness, H, was then determined as described above. We note however, that this calculated crustal thickness represents the average for the region, and not the zero-elevation reference crustal thickness H_0. This later value was determined by assuming the average elevation, $\bar{h}$, of the region was compensated by an Airy mechanism using the equation

$$H = H_0 + \bar{h}\left[1 + \frac{\rho_c}{\rho_m - \rho_c}\left(\frac{R_0}{R_0 - H}\right)^2\right]. \tag{8}$$

We assume here that the degree-1 topography of Mars is compensated by crustal thickness variations, and thus have included the degree-1 topography terms when computing $\bar{h}$. To a good approximation, $\bar{h}$ simply represents the degree-1 topography for each region.

[46] The results from this synthetic test are displayed in Figure 2 as histograms of the zero-elevation crustal thickness H_0 and the x-intercept h_0. Specifically, results are shown for the cases where geoid and topography data were regressed within circles of radii 500, 1000, 1500, and 2000 km. As is seen, the recovered value of H_0 is within a kilometer of the true value when the circle radius is greater or equal to 1000 km. In addition the uncertainty in this value is seen to decrease slightly with increasing circle radii. The value of h_0 should be zero when the degree-1 topography has been removed before determining the geoid-to-topography ratio. As is seen in Figure 2, this expectation is recovered when the circle radii are greater or equal to 1500 km. If noise was added to the geoid data before regressing for the GTR, a larger circle radius would be required to reduce the uncertainty in the crustal thickness estimates. On the basis of these synthetic results, we chose to use a circle radius of 2000 km when analyzing the real Martian data.

4. Results

[47] The main benefit of using a spatial GTR analysis over that of a spectral-admittance analysis in interpreting gravity and topography data is that it is possible to easily ignore regions that do not conform to the expectations of the assumed compensation model. For example, if one were to calculate spectral admittances for the southern highlands using a Cartesian Fourier analysis, then it would be necessary to restrict the analysis to a square region that was not influenced by any of the mascon basins, the south-polar ice cap, nor any of the numerous volcanic constructs that are located there. In addition, the size of such a box should be considerably greater than the spectral resolution inherent in the utilized gravity field (i.e., spherical harmonic degree $\sim$60, or a wavelength of $\sim$350 km which corresponds to $\sim 6^\circ$ of latitude) in order to obtain numerous estimates of the wavelength-dependent admittance function. Those regions of the highlands that are thus amenable to such a spectral analysis are severely restricted by these considerations.

[48] In this analysis, GTRs were only computed and interpreted for those regions of the southern highlands that are likely to be fully compensated. As was discussed in section 2, the assumption of isostasy is expected to be a good approximation for most of the southern highlands [*Phillips and Saunders*, 1975; *Frey et al.*, 1996; *McGovern et al.*, 2002]. Those regions of the Martian surface that are likely to be at least partially supported by the strength of the lithosphere, or may possess significant lateral variations in density, were simply ignored. Specifically, we have conservatively excluded the following regions and geologic provinces from our analysis: (1) the Tharsis plateau, (2) the Argyre and Isidis mascon basins, (3) the south-polar ice cap, (4) the northern lowlands, and (5) various highland volcanic constructs, including Syrtis Major and Hesperia Planum, and Hadriaca, Tyrrhena, Apollinarus, Peneus and Amphitrites Paterae. The resulting study region is delineated by the white lines in Figures 4 and 5.

[49] An additional non-isostatic feature of Mars that is more challenging to remove is the global gravity and

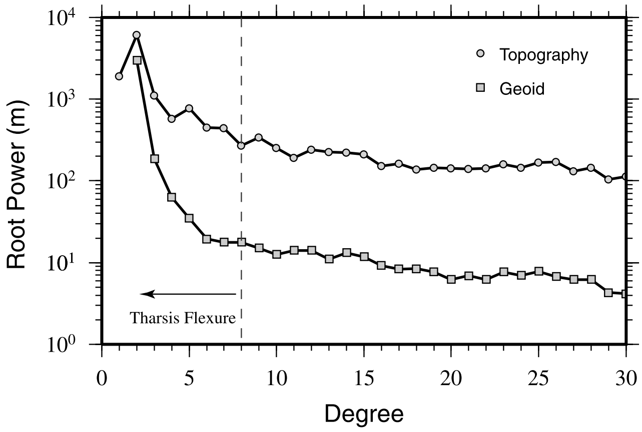

Figure 3. The square root of the power for the geoid and topography plotted as a function of spherical harmonic degree. The high power associated with the low-order degrees is most likely a result of the load and global flexure associated with the Tharsis province [e.g., *Zuber and Smith*, 1997; *Phillips et al.*, 2001].

topography signal associated with the load and flexure of the Tharsis province. By treating the Martian lithosphere as elastic, and taking into account both bending and membrane stresses [e.g., *Turcotte et al.*, 1981], *Phillips et al.* [2001] showed that the massive volcanic load associated with the Tharsis plateau should significantly influence both the long-wavelength gravity and topography of this planet. The response of this load is most simply described as a gravity and topographic low surrounding the Tharsis province (the "Tharsis trough"), and an antipodal gravity and topographic high near Arabia Terra (just north of the Hellas basin). As this global signal cannot be excluded in the space domain simply by ignoring the Tharsis province, we have attempted to minimize this signal by spectral filtering.

[50] In Figure 3 the square root of the power associated with the topography (Mars2000.shape [*Smith et al.*, 2001]) and geoid (mgm1025 [*Lemoine et al.*, 2001]) is plotted as a function of spherical harmonic degree. The power in both the geoid and topography for degrees less than ~6 is seen to be significantly greater than would be expected from a simple extrapolation of the higher degree data. As modeled by *Phillips et al.* [2001], the deformation of Mars due to the Tharsis load results primarily from membrane stresses and is hence only important for the lowest-degree spherical harmonics (in their study, they modeled the Tharsis load up to degree 10). Hence the anomalously high power in this figure for degrees less than ~6 is most easily interpreted as resulting from this global deformation. In a different study, *Zuber and Smith* [1997] modeled the gravitational attraction that is associated with the Tharsis province and found that it in fact dominates the Martian gravity field for degrees less than ~6.

[51] While one could attempt to "remove" the Tharsis signature from the low-degree gravity and topography terms, such a process would depend on many assumptions. If one were to assume that the lithosphere behaved elastically over the past 4 Ga, then one would need to know the lithospheric thickness and the magnitude and geometry of the initial load. Instead, if one were to assume that some of the elastic stresses have relaxed by viscous processes, then one would need to model the temperature and strain rate dependent viscosity. As these quantities are likely to never be perfectly known, we chose instead to remove the lowest degree terms in this analysis and investigate the consequences of truncating various portions of the low degree spherical harmonic terms. While we acknowledge that truncation is not perfect as the "highland" and "Tharsis" signals are likely to overlap in the wavelength domain, imperfect modeling of the Tharsis flexural signal would likely give rise to substantial error in the GTR analysis as the lowest-degree spherical harmonic terms heavily influence the calculated GTR.

[52] Before presenting our main results, we first demonstrate the consequences of retaining the Tharsis signal in a GTR analysis. For this scenario, we have assumed a crustal density of 2900 kg m^{-3}, and have removed the degree-2 gravity and topography terms. The computed zero-elevation crustal thicknesses for the southern highlands are displayed in Figure 4, and it is seen that they range from less than zero, to more than 400 km. A similar, though worse, result is obtained if only the J_2 terms which are dominated by the rotational flattening are removed. Also shown in this figure is the single profile that was used in the GTR analysis of *Turcotte et al.* [2002] to infer a crustal thickness of 90 ± 10 km. While our results for the same region are in accord with their value, it is clear that no single crustal thickness can be taken as being representative of Mars from this analysis. The reason for this is clear from equation (2) and Figure 3: GTRs are most heavily weighted by those admittances, Z_l, that possess the greatest topographic power, and these degrees are dominated by the flexural response of the Tharsis plateau.

[53] In Figure 5, we plot the calculated crustal thickness for the case where we have truncated the low-degree gravity and topography terms that are less than or equal to degree 10. The average uncertainty associated with each individual crustal thickness determination is approximately ±3 km. In comparison with the previous figure, the highlands are now found to possess a relatively constant zero-elevation crustal thickness of about 60 km, with values greater than 100 km and less than 40 km being nearly absent. (We note that the actual crustal thickness for each region could be computed using equation (8).) The consequences of truncating the gravity and topography fields at different degrees are illustrated in Figure 6 where we plot histograms of the zero-elevation crustal thickness and h_0 for the cases where only degrees greater than 5, 7, 9 and 10 were utilized. With regard to the histograms for h_0, it is seen that the distributions are considerably broader when the gravity and topography are truncated at the lower degrees. This is most easily explained if the degrees between 5 and 7 still possess some amount of "Tharsis contamination." A similar phenomenon is observed in the histograms of the zero-elevation crustal thickness.

[54] On the basis of the theoretical expectation that h_0, should be zero, as well as the likely possibility that a non-negligible Tharsis flexural signal might still be present in the degrees between 6 and 9, we consider crustal thickness estimates only for the cases when degrees greater than 9 and 10 were used in the gravity and topography expansions. We have investigated the effects of truncating these fields at

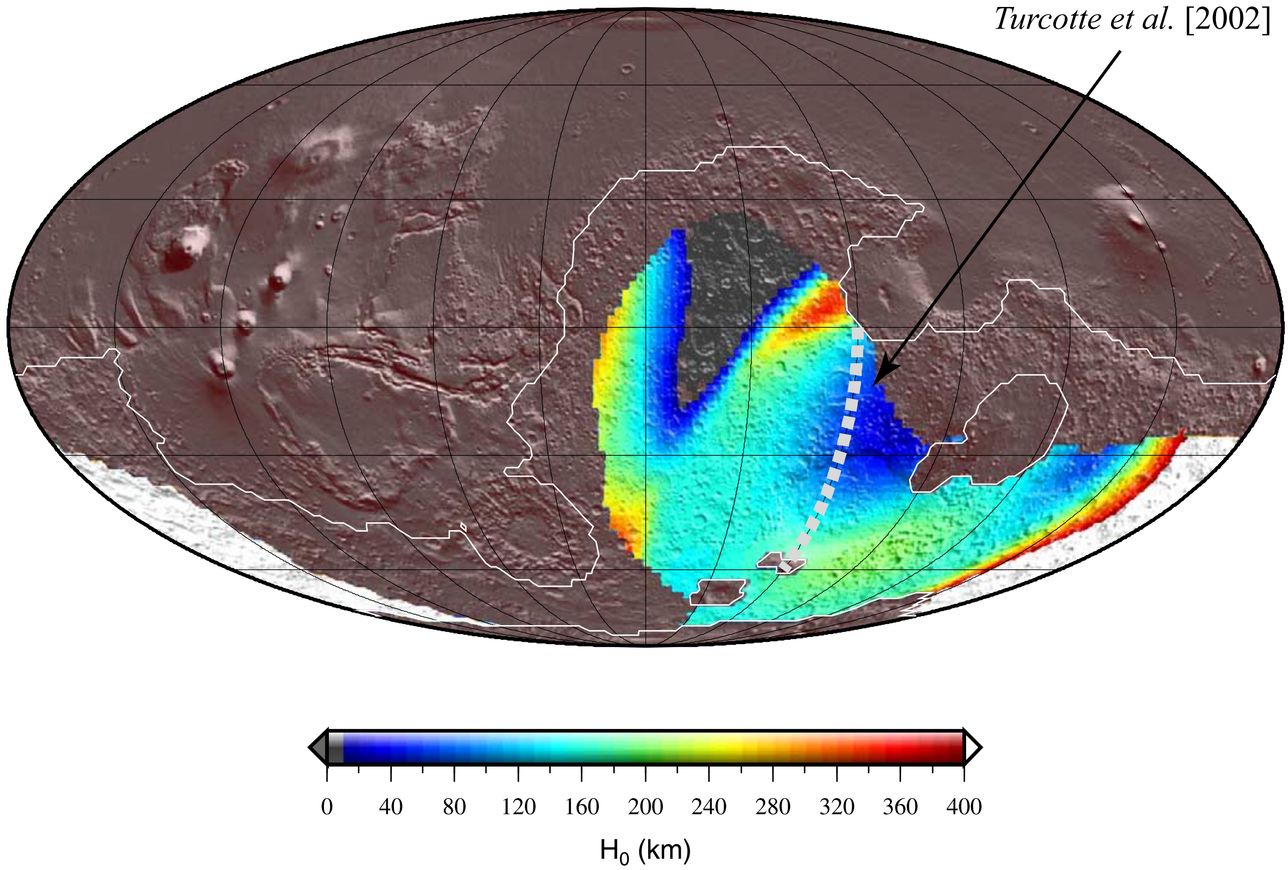

Figure 4. Calculated zero-elevation crustal thicknesses for the southern highlands for the case where only the degree-2 terms of the geoid and topography were removed. The thin white lines delineate those regions which were used in computing the GTRs. The heavy dotted line corresponds to the profile analyzed by *Turcotte et al.* [2002] where a crustal thickness of 90 ± 10 km was obtained. The color scale is saturated for crustal thicknesses greater and less than 400 and 0 km, respectively.

higher degrees, and have found that the 1-σ crustal-thickness limits do not change significantly until the point where only degrees greater than 15 to 20 where retained in the analysis, at which point the average crustal thickness decreased by about 10 to 15 km. We regard this latter effect as a likely consequence of discarding the "highland signal" with progressive amounts of truncation.

[55] As neither the density nor composition of the highland crust is well constrained, we have also employed a conservative range of crustal densities between 2700 and 3100 kg m^{-3} when inverting for crustal thickness. The lower bound of this range is typical of terrestrial crustal rocks, and is in agreement with the admittance study of Noachis Terra performed by *McGovern et al.* [2002], whereas the upper bound is consistent with the densities of the Martian volcanoes as determined from the admittance studies of *McGovern et al.* [2002] and *McKenzie et al.* [2002]. While we find the unfractured densities of the Martian basaltic meteorites to be greater than this range (~3300 kg m^{-3}), and those of the cumulate rocks to be even greater (~3500 kg m^{-3}), it is most likely that these rocks have an origin within the Tharsis plateau and/or Elysium volcanic complex [e.g., *McSween*, 1985, 1994] and are not representative of those rocks that comprise the bulk of the ancient southern highlands.

[56] We list our final results in Table 2 for two different truncation-degrees and three plausible crustal densities. If we take the 1-σ limits of all the estimates in this table, we find the zero-elevation crustal thickness of the highlands to lie between 33 and 81 km (or 57 ± 24 km). We note, however, that if the crustal density was known a priori, that it would then be possible to decrease this range of values by about 15 km.

[57] Finally, we note that there is a single apparently anomalous region just northwest of the Hellas basin that possesses higher than typical zero-elevation crustal thicknesses. As this region does not correlate with any specific geologic feature, the most likely explanation is that either the assumption of Airy isostasy is inappropriate there, or that the process of truncating the low-degree gravity and topography terms has somehow affected this region (either by discarding the highland signal, or by incomplete removal of the Tharsis signal). If this anomaly was solely a result of the crust there having a higher than typical density, a density of at least 4500 kg m^{-3} would be required which is higher than any known Martian sample. While dense intrusive rocks could conceivably account for a portion of this anomaly, the lack of surface volcanic features argues against such an interpretation. We have tested whether this anomaly might be caused by the Hellas basin itself, but ignoring this

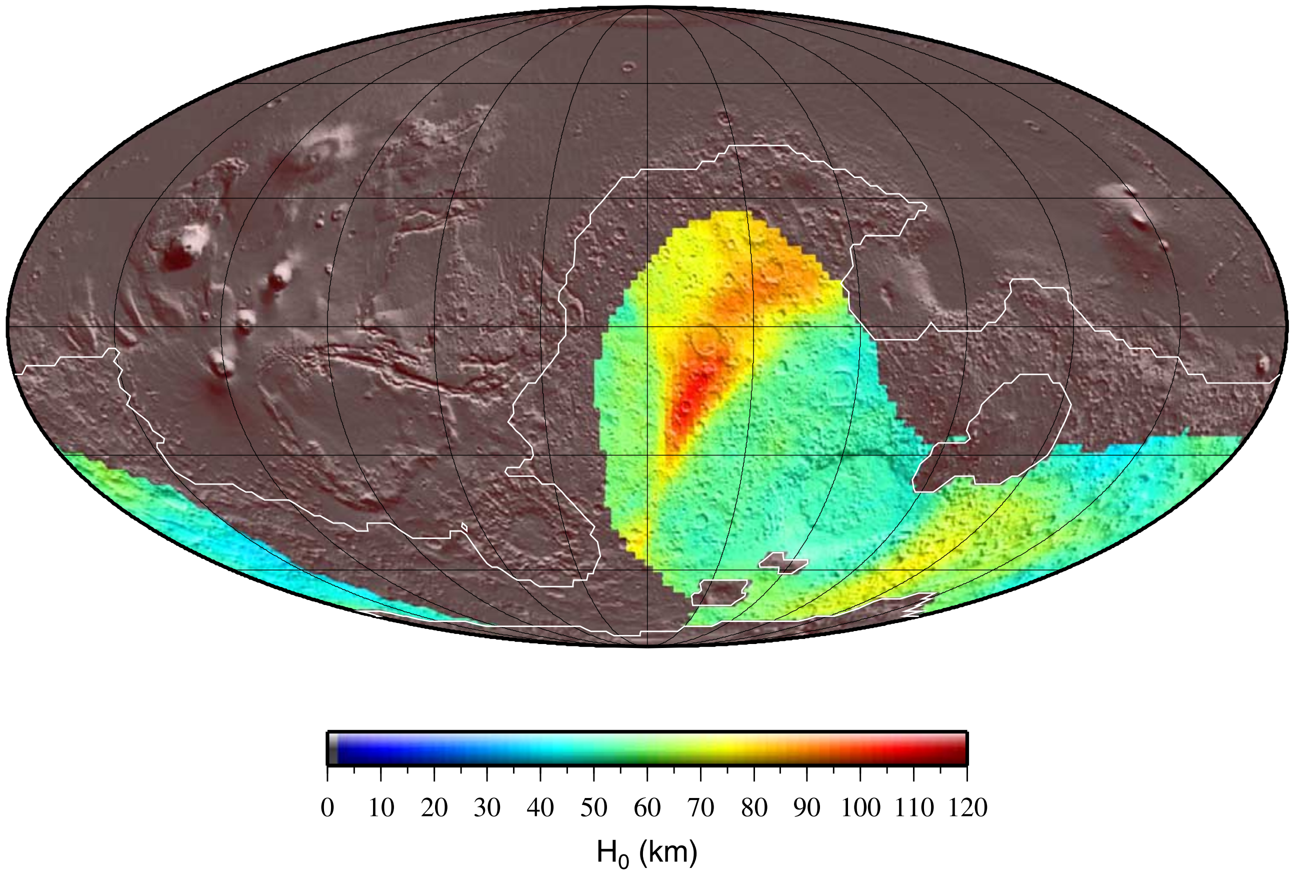

Figure 5. Calculated zero-elevation crustal thicknesses for the southern highlands for the case where degrees less than or equal to 10 were removed from the geoid and topography. Geoid and topography data were regressed within circle radii of 2000 km, and a crustal density of 2900 kg/m^3 was assumed.

basin in the GTR analysis yields similar results. As this region is relatively small, it does not significantly affect the average crustal thickness of the southern highlands, but does give rise to a high crustal-thickness "tail" in the histograms displayed in Figure 6.

[58] The crustal thickness estimates obtained in this study are consistent with the most reliable previous estimates listed in Table 1. In particular, our lower bound of 33 km is greater than the minimum allowable value of 32 km based on global crustal thickness considerations, and our range overlaps that obtained from the spectral-admittance study of *McGovern et al.* [2002] (6–68 km). Furthermore, our maximum value of 81 km is less than the maximum allowable value of 115 km based on the viscous relaxation study of *Nimmo and Stevenson* [2001], as well as the 93-km maximum value obtained from thorium mass balance considerations. If all of the estimates listed in Table 1 are equally accurate, then a crustal thickness range from 38 to 62 km (or 50 ± 12 km) is consistent with all studies.

5. Implications for the Thermal Evolution of Mars and the Origin of its Crust

[59] The growth of planetary crusts is a complicated process that involves both "primary" differentiation events at the time of accretion and subsequent "secondary" partial melting of the mantle later in time. In the case of the Earth, reprocessing of mantle-derived crustal rocks has further given rise to a "tertiary" crust which appears to be unique in the solar system [e.g., *Taylor*, 1992]. In this section, we briefly discuss the first-order implications that the thickness of the Martian crust has for the thermal evolution of this planet and the origin of its crust. For comparative purposes, we first briefly summarize how the thickness of the lunar and terrestrial crusts are related to the thermal evolution of these bodies.

[60] The time-integrated thermal history of the Moon is more straightforward than that of the Earth and Mars. The vast majority of the lunar crust is believed to have formed during a short period of time (~100 Ma) by the flotation of anorthositic mineral assemblages in a near-global magma ocean [e.g., *Warren*, 1985; *Pritchard and Stevenson*, 2000]. Minor additions to this primary crust occurred from partial melting of the lunar interior over a period of about 3 Ga [e.g., *Head and Wilson*, 1992; *Hiesinger et al.*, 2000]. A first-order indicator of the efficiency and importance of crustal production is the percentage of the silicate portion of a planet that is crustal in origin. Assuming a lunar core radius between 0 and 375 km (upper bound taken from *Williams et al.* [2001]), and a crustal thickness range between 43 and 53 km (using the seismic constraints of *Khan and Mosegaard* [2002] and the methodology of *Wieczorek and Phillips* [1998]), the lunar crust represents between 7 and 9% of its silicate volume. Estimates for the

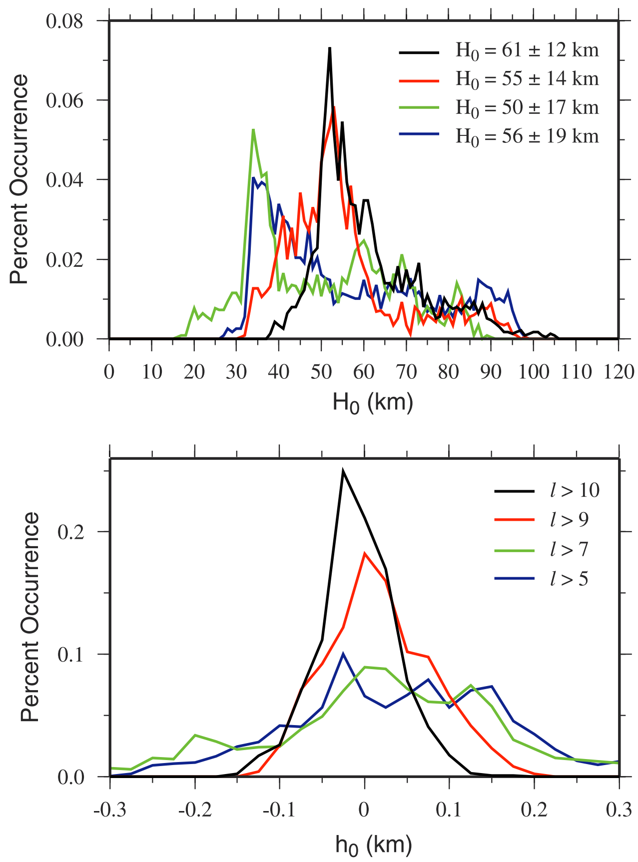

Figure 6. Example histograms of the calculated zero-elevation crustal thickness H_0 and x-intercept h_0. Shown are the cases where only degrees greater than 5, 7, 9 and 10 were used in the calculations. Circle radii of 2000 km were used to regress geoid and topography data and a crustal density of 2900 kg m^{-3} was assumed.

volume of mantle-derived extrusive mare basalts are generally about 1% of this crustal volume [e.g., *Head and Wilson*, 1992], with the intrusive equivalent of these basalts comprising an additional few percent [*Wieczorek and Zuber*, 2001]. Since the secondary crustal component is small, the crustal thickness of the Moon is principally controlled by primary differentiation processes that include the depth extent of the magma ocean and how efficiently buoyant anorthositic mineral assemblages were sequestered into the crust.

[61] In contrast to the Moon, the Earth represents the opposite extreme in possessing only a secondary and tertiary crust, its primary crust having been recycled early in Earth history. Most of the secondary crust of the Earth is formed by partial melting of the mantle beneath oceanic ridges, with minor additions arising from continental flood basaltic volcanism. Factors which influence the thickness of the oceanic crust include the spreading rate, and the manner in which melt is focused beneath a ridge via porous flow [e.g., *Sparks and Parmentier*, 1994; *Kelemen et al.*, 1999]. While the continental crust is composed of accreted oceanic crust, subduction-related volcanics, and continental flood basalts, there is no unified theory that can predict its average thickness. Assuming that 63% of the Earth's surface is oceanic with a crustal thickness of 7 km, and that the remainder is continental crust with a mean thickness of 35 km, only 1% of the silicate portion of the Earth is composed of crustal materials. At present, only about 25% of this crustal volume is oceanic. Nevertheless, as oceanic crust is continually being created and recycled, the total amount of crust that was produced over the Earth's entire history should be much larger than this. Extrapolating current oceanic crustal production rates backward in time suggests that the total amount of oceanic crust produced is about 20 times more than the present-day value [e.g., *Taylor*, 1992].

[62] The crustal evolution of Mars is likely to lie somewhere between the extremes of the Earth and Moon. Assuming a crustal thickness of 38 to 62 km, and a core radius from 1150 to 1850 km [*Harder*, 1998; *Yoder et al.*, 2003], between 3 and 6% of the silicate portion of Mars is currently composed of crustal materials. This corresponds to about half of that of the Moon, suggesting that crustal production was comparatively less efficient on this body, possibly as a result of slower accretion timescales. However, if any form of crustal recycling occurred on Mars (as does on the Earth), then the total amount of crust produced over Martian history could have been considerably larger. One manner in which the crust could have been recycled into the mantle is by the delamination of dense eclogitic lower crustal materials. However, as the eclogite phase transition is predicted to occur about ~150–200 km below the surface [*Sohl and Spohn*, 1997], and as the Martian crustal thickness is likely to be everywhere thinner than this [*Zuber et al.*, 2000], it is unlikely that this form of crustal recycling has ever occurred. Alternatively, it has been suggested that an early period of plate tectonics may have operated in the northern lowlands of Mars [*Sleep*, 1994]. While such a scenario could plausibly lead to a core dynamo and explain the ancient crustal magnetic anomalies of this planet [*Nimmo and Stevenson*, 2000], plate tectonics would also efficiently cool the mantle and give rise to a thinner than observed crust [*Breuer and Spohn*, 2003]. Isotopic considerations, as discussed below, additionally argue against early crustal recycling. Furthermore, while the surface age of the northern lowlands is younger than that of the southern highlands, it has been shown by *Frey et al.* [2002] that the *basement* of the northern lowlands (which underlies a thin volcanic or sedimentary cover) is in fact at least as old as the visible southern highlands. Such an observation is difficult to reconcile with the early operation of plate tectonics in the northern lowlands as originally envisioned by *Sleep* [1994].

[63] Isotopic systematics of the Martian meteorites suggest that Mars underwent a major differentiation event early in its history. In particular, ^{146}Sm-^{142}Nd isotopic data (with

Table 2. Crustal Thickness Limits as Determined From a GTR Analysis of the Southern Highlands of Mars

	H_0, km			
ρ_c, kg m^{-3}	2700	2900	3100	1-σ Limits
l > 10	68.3 ± 12.7	60.7 ± 12.4	52.8 ± 13.8	39–81
l > 9	61.6 ± 14.6	54.7 ± 13.7	47.4 ± 14.3	33.1–76.2

a half-life of 103 Ma) indicate that this planet underwent a major silicate differentiation event between 0 and 33 Ma [*Harper et al.*, 1995; *Borg et al.*, 1997], whereas Hf-W isotopes (half-life of 9 Ma) indicate that Martian core formation occurred within about 30 Ma [*Lee and Halliday*, 1997]. Indeed, the correlation between these two isotopic systems highly suggest that core formation and silicate differentiation were contemporaneous over this time period [*Lee and Halliday*, 1997; *Blichert-Toft et al.*, 1999]. The existence of ε_W anomalies in the Martian meteorites argues for an isotopically heterogeneous mantle that has not undergone convective homogenization [*Lee and Halliday*, 1997], and the large positive $\varepsilon^{142}{}_{Nd}$ anomalies present in young mantle-derived rocks argues against significant crustal recycling having affected their source region [*Borg et al.*, 1997]. Indeed, in considering both Rb-Sr and Sm-Nd isotopic systematics, *Borg et al.* [1997] concluded that "The apparent absence of crustal recycling and the relative isolation of the Martian mantle from the Martian crust are the most significant geologic differences between Mars and the Earth."

[64] These isotopic considerations suggest that a primary crust was produced early in Martian history (~30 Ma), and furthermore that is has survived and remained isotopically isolated from the mantle to the present-day. The geochemically based crustal thickness estimate of *Norman* [1999] based on a Nd mass balance calculation (see section 2.8) is most applicable to the thickness of this primary crust. If his results are taken at face value (with a best-fit primary crustal thickness lying between 20 and 30 km), then at least one third, and possibly the vast majority, of the present-day Martian crust is primary in origin. As of yet, though, no Martian meteorites have been identified that might represent this vast geochemical crustal component [e.g., *Lee and Halliday*, 1997; *Münker et al.*, 2003]. Obtaining such a sample (presumably in the Martian southern highlands) should thus be considered as a primary objective of any Martian sample return mission. (With the possible exception of ALH84001, all Martian meteorites are believed to have come from either the young Tharsis province or Elysium Mons [*McSween*, 1994; *Head et al.*, 2002]. Though ALH84001 has an ancient age, its isotopic composition is inconsistent with being a sample of this hypothetical "enriched crust.")

[65] The total crustal thickness of a planet is directly related to its time-integrated magmatic evolution and hence should be satisfied by any model of its thermal evolution. Nevertheless, while a number of Martian thermal models have been published, very few have calculated the volume of crust that is produced, and of those that did, many did not take into account the latent-heat of melting nor reasonable melting relationships. A notable exception is the study of *Hauck and Phillips* [2002], in which the thermal evolution of Mars was modeled by coupling lithospheric growth, melt production, the fractionation of heat-producing elements into the crust, and parameterized mantle convection. A successful model was defined as one that produced a crustal thickness between 50 and 100 km and that mostly formed within the first 500 Ma of Martian history (this latter constraint being based on the observation that the orientation of Noachian-aged valley networks are controlled by the flexure associated with the Tharsis plateau [*Phillips et al.*, 2001]). The effects of the assumed input parameters were then systematically investigated to see how they affected the modeled crustal growth-rate curves.

[66] One of the major results of that study was to show how the total amount of crust produced was related to the assumed bulk abundances of heat-producing elements. In particular, those models that incorporated bulk potassium abundances greater than chondritic [e.g., *Schuber and Spohn*, 1990; *Lodders and Fegley*, 1997] were found to produce a crust about 250 km thick, or approximately four times greater than the maximum crustal thickness of 62 km as advocated here. In order to satisfy the constraint that most of the crust should form in the first 500 Ma of Martian history, they further required a mantle rheology typical of a wet olivine flow law (a dry flow law caused the bulk of crustal production to occur after ~3 Ga). It was also shown that if Mars possessed a primary crust that was enriched in heat-producing elements that the amount of subsequent crustal production would be reduced. As discussed above, such a primary crust likely comprises a large portion of the present-day crust of Mars, suggesting that subsequent crustal production may have been relatively unimportant (and possibly largely confined to the Tharsis province).

6. Conclusions

[67] We have used geoid-to-topography ratios (GTRs) to place constraints on the thickness of the southern-highland crust using the spectrally weighted spherical admittance approach of *Wieczorek and Phillips* [1997]. We have demonstrated that (1) a Cartesian approach underestimates the crustal thickness by tens of kilometers and (2) one must first remove the global loading and flexural signal associated with the Tharsis province before performing such an analysis. Assuming a conservative range of crustal densities (2700–3100 kg m^{-3}) the average crustal thickness is constrained to lie between 33 and 81 km (or 57 ± 24 km).

[68] While a large number of crustal thickness estimates have been published in the literature, a critical review leads us to advocate only a few. In particular, viscous relaxation studies of *Zuber et al.* [2000] and *Nimmo and Stevenson* [2001] place defensible upper bounds on the average crustal thickness of ~100 km, while a thorium mass balance calculation (based on the study of *McLennan* [2001]) implies an upper bound of 93 km. A spectral-admittance study of highland features by *McGovern et al.* [2002] constrains the average thickness of the crust to lie between 8 and 68 km (with a best fit value of 50 km), and by requiring the thickness of the crust to be everywhere greater than 3 km, global crustal thickness modeling places a lower bound of 32 km on the average crustal thickness. If these estimates are all equally accurate (see Table 1), then a Martian crustal thickness between 38 and 62 km (or 50 ± 12 km) is consistent with all studies.

[69] Neodymium and tungsten isotopic constraints imply that Mars underwent a major differentiation event during the first 30 Ma of its evolution, creating distinct crustal and mantle isotopic reservoirs [*Harper et al.*, 1995; *Borg et al.*, 1997; *Lee and Halliday*, 1997; *Blichert-Toft et al.*, 1999]. The preservation of these isotopic anomalies in young mantle-derived partial melts implies that crustal recycling has not since occurred. A neodymium mass balance study

[*Norman*, 1999] further suggests that the thickness of this primary crust is between 20 and 30 km thick, corresponding to more than one third of the present-day crustal thickness. This primary enriched crust has not yet been directly identified in the presently known Martian meteorites.

[70] The average Martian crustal thickness is an important constraint for models of thermal evolution. One recent study by *Hauck and Phillips* [2002] has systematically investigated the growth of the Martian crust, and has determined that: (1) bulk compositional models which employ potassium abundances greater than chondritic give rise to greater than admissible crustal thicknesses, and (2) a wet olivine flow law is required for the mantle in order to produce most of the Martian crust before 4 Ga. These conclusions are strengthened by the crustal thickness estimates of this study which are slightly less than those employed by *Hauck and Phillips* [2002]. They have further shown that a primordial crust enriched in heat-producing elements would tend to decrease the amount of subsequent melting in the mantle. As a primordial crust is predicted to comprise a large fraction of the present-day Martian crust, this effect should not be neglected in future thermal evolution models [cf. *Parmentier and Zuber*, 2001].

[71] Finally, we note that even though geophysical studies appear to be converging on an average Martian crustal thickness of ~50 km, significant uncertainties warrant future studies utilizing different and more sophisticated techniques. First and foremost among these is an almost complete lack of knowledge of the bulk crustal composition, and hence its density. Most studies assume that these properties are uniform beneath the southern highlands, but if lessons learned from lunar studies have any relevance [e.g., *Jolliff et al.*, 2000], this is surely an oversimplification. Future seismic studies of Mars [e.g., *Lognonné et al.*, 2000] would help resolve these issues, but unfortunately there are currently no plans for such a mission.

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