

Evidence for a differentiated crust in Solis Planum, Mars, from lithospheric strength and heat flow

Javier Ruiz ^{a,*}, Rosa Tejero ^a, Patrick J. McGovern ^b

^a *Departamento de Geodinámica, Facultad de Ciencias Geológicas, Universidad Complutense de Madrid, 28040 Madrid, Spain*

^b *Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston, TX 77058, USA*

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Abstract

Two independent sets of heat flow estimates provide constraints on the Hesperian-era surface and mantle heat flows, and the thickness of the heat-producing elements (HPE)-enriched upper crust, in the Solis Planum region of Mars. The calculations, which use the concentration of uppermost crust heat sources deduced from orbital gamma ray spectroscopy and soils geochemistry, are based on the effective elastic thickness of the lithosphere and the minimum depth of faults underlying winkle ridges. We find that, for the majority of analyzed settings, the HPE-enriched crust is thinner than the whole crust thickness in this region (~ 65 km). Thus, our results strongly support a differentiated martian crust.

Keywords: Mars; Geophysics; Tectonics

1. Introduction

The thermal structures of planetary lithospheres have been constrained by their mechanical responses to loading or faulting, as inferred from remotely sensed data. Surface heat flows estimates for diverse places on Mars have been derived from the effective elastic thickness of the lithosphere (e.g., Solomon and Head, 1990; Anderson and Grimm, 1998; Zuber et al., 2000; McGovern et al., 2002, 2004; Ruiz et al., 2004, 2005) or from faulting depth (Schultz and Watters, 2001). Similarly, a model of winkle ridge origin by localization instability has been used to constrain thermal gradients on martian ridged plains (Montesi and Zuber, 2003). These calculations are valid for the time when the structures used as indicators were formed. However, the need to consider the (previously ignored) presence of heat sources within the martian crust to calculate surface heat flow and temperature-depth profiles from elastic thicknesses has been recently pointed out (Ruiz et al., 2004): indeed, greater crustal heat production implies higher surface heat flow and lower temperatures at the base of the crust.

The analysis of martian soils and orbital gamma ray spectroscopy shows significant amounts of heat-producing elements (HPE) on the martian surface; such measurements are considered as representative of, at least, the upper crust (McLennan, 2001, 2003; Taylor et al., 2003a). There is no reason for which the HPE-enriched crust should coincide with the whole crust. Indeed, the martian crust could be stratified. Moreover, whereas a combination of several geophysical arguments constrains the average crustal thickness to between 38 and 62 km (Wieczorek and Zuber, 2004), a light rare Earth elements (LREE)-enriched crust (again not necessarily equivalent to the whole crust) ≤ 45 km thick, and most probably 20–30 km thick, has been proposed on the basis of the geochemistry of martian meteorites (Norman, 1999).

For the case of Solis Planum region, two different set of observations can be used as heat flow indicators for the Hesperian epoch: the effective elastic thickness and the depth of faults underlying winkle ridges. Results obtained from both these heat flow indicators should be roughly consistent. For that reason, these results are used in this work in order to establish constraints to the thermal properties of the lithosphere in Solis Planum at the Hesperian, including surface heat flow, crust and

* Corresponding author.
E-mail address: jaruiz@geo.ucm.es (J. Ruiz).

mantle contribution to the total heat flow, and the thickness of the HPE-enriched crust.

2. Temperature profiles

The whole crust thickness in Solis Planum is approximately 65 km according to Neumann et al. (2004) (see their Fig. 7); we adopt this value here. Temperature profiles are calculated by assuming heat sources homogeneously distributed in a HPE-enriched upper crust of thickness b (between 0 and 65 km), and linear thermal gradients for the lower part of the crust (with thickness equal to 65 km $- b$) and the upper mantle. Hence, in the HPE-enriched crust, temperature at a depth z is

$$T_z = T_s + \frac{Fz}{k} - \frac{Az^2}{2k}, \quad (1)$$

where T_s is the surface temperature, F is the surface heat flow, k is the thermal conductivity of the crust, and A is the volumetric heat production rate. Linear temperature gradients in lower crust and upper mantle are calculated for a constant mantle heat flow (in essence the heat flow from the convective interior) given by $F - Ab$. Whereas a linear gradient would be a useful approximation for the upper mantle, the distribution of HPE in a putative martian lower crust is unknown. On Earth, radiogenic sources are sparse beneath the upper crust, but it is not necessarily true for Mars. In any case, a thermal gradient should be roughly adequate for a HPE-poor crust.

Surface temperature is taken as 220 K, the present-day mean surface temperature (Kieffer et al., 1977). Thermal conductivity is assumed as $2.5 \text{ W m}^{-1} \text{ K}^{-1}$ for both upper and lower crust, which is the mean value for diabase (Kobranova, 1989), a rock type commonly taken to represent the mechanical behavior of the martian crust, and it has been frequently used for the martian crust (Solomon and Head, 1990; McGovern et al., 2002, 2004). For the lithospheric mantle thermal conductivity is assumed as $3.5 \text{ W m}^{-1} \text{ K}^{-1}$ (e.g., Burov and Diament, 1995).

The volumetric heating rate in the upper crust depends on time before present and the amount of HPE, essentially potassium, thorium and uranium. In the calculations we use a local crust density of 2900 kg m^{-3} (McGovern et al., 2002, 2004), decay constants from Turcotte and Schubert (2002), and the maximum possible age range for the Hesperian period, 2.9–3.7 Gyr (Hartmann and Neukum, 2001). Potassium and thorium abundances, as well as K/Th ratios show a wide variation on the surface (Taylor et al., 2003a). K and Th abundances at Solis Planum are relatively low, but we use average values for the HPE-enriched crust: although Solis Planum region is covered with Hesperian volcanic materials (e.g., Tanaka et al., 1992), the bulk of the martian crust was in place by the Early Noachian (e.g., Solomon et al., 2005a). The average concentration of thorium observed by the Mars Odyssey GRS is 1.1 ppm (Taylor et al., 2003a). Assuming this value as representative for the (at least) upper crust, and the Th/U ratio as in McLennan (2001, 2003), gives an uranium concentration of 0.3 ppm. If potassium concentration is taken as 0.35% (Taylor et al., 2003a, 2003b; see also the Mars Odyssey GRS web site at <http://grs.lpl.arizona.edu>), then $A \approx 0.5 \text{ } \mu\text{W m}^{-3}$ for 2.9 Gyr

ago, and $A \approx 0.7 \text{ } \mu\text{W m}^{-3}$ for 3.7 Gyr ago. These values are similar to those obtained from K, Th and U abundances in McLennan (2001), and lower than the deduced from McLennan (2003), works which were based on Viking and Pathfinder soils and Phobos 2 GRS. We also consider the effect of varying A in the calculations.

3. Heat flows from effective elastic thickness

Estimates of effective elastic thickness from gravity/topography admittance and correlation spectra fall in the range 24–37 km for Solis Planum (McGovern et al., 2002, 2004). These values correspond to (McGovern et al., 2004) linear thermal gradients of $8\text{--}14 \text{ K km}^{-1}$ and heat flows (for a crust thermal conductivity of $2.5 \text{ W m}^{-1} \text{ K}^{-1}$) of $20\text{--}35 \text{ mW m}^{-2}$, in the absence of heat sources within the crust. Here we use these elastic thickness values in order to perform a finer calculation of heat flows and lithospheric temperatures by including crustal heat sources.

The effective elastic thickness is a measure of the total strength of the lithosphere, which integrate contributions from brittle and ductile layers and the elastic core of the lithosphere (for reviews see Watts, 2001; Watts and Burov, 2003). Effective elastic thicknesses are converted to heat flows and temperature profiles following the equivalent strength envelope procedure described by McNutt (1984), as adapted by Ruiz et al. (2004, 2005) to take into account heat sources in an upper crust equal to or thinner than the whole crust. This methodology is based in that the bending moment of the mechanical lithosphere is equal to the bending moment of the equivalent elastic layer of thickness T_e . The bending moment of the mechanical lithosphere is estimated from its strength envelope and from the curvature of the elastic layer: the link to heat flow comes from the dependence of the ductile strengths on the temperature profile.

Heat flow calculations for Solis Planum by McGovern et al. (2002, 2004) only consider crustal strength. The base of the mechanical lithosphere can be defined as the depth at which the strength reaches a low value, usually taken to be in the range 10–50 MPa (McNutt, 1984; Ranalli, 1994; Anderson and Grimm, 1998), with the lower value being more appropriate for the low gravity of Mars (Ruiz et al., 2005). Geotherms and strain rates from McGovern et al. (2004) give significant ductile strength at the top of the mantle for Solis Planum at the Hesperian, 15–9400 MPa for a dry olivine rheology (Chopra and Paterson, 1984). Whereas wet olivine might be appropriate for the deep and convective mantle of Mars (Hauck and Phillips, 2002; Solomon et al., 2005b), dry olivine is more probable for the uppermost mantle on the basis that melting, related to crust extraction from upper mantle, is very effective in water extraction (Montesi and Zuber, 2003). In a similar way, a dry rheology has been usually considered for the oceanic mantle lithosphere on the Earth (e.g., Kohlstedt et al., 1995), precisely due to the extraction of the basaltic crust in these regions. Thus, this implies that the strength of the martian mantle should be not neglected in the calculations here presented. Taking into account lithospheric mantle strength increases total strength of the

lithosphere, which must be compensated with higher heat flow than in calculations that consider only crustal strength.

Again for geotherms and strain rates in [McGovern et al. \(2004\)](#), ductile strength at the base of the crust for Solis Planum at the Hesperian is 0.1–4 MPa, assuming a diabase rheology ([Caristan, 1982](#)) which should be adequate for a basaltic crust. These values are lower than those used to define the base of the mechanical lithosphere, which implies mechanically decoupled crust and mantle. We therefore modified the equivalent strength envelope procedure so that it is applicable to a rheologically stratified lithosphere, estimating the total bending moment of the lithosphere from the respective contributions of crust and mantle. Taking equal elastic constants and curvature values for the crust and upper mantle, the total bending moment of a lithosphere with mechanically decoupled crust and mantle is ([McNutt et al., 1988](#); [Burov and Diamant, 1992](#))

$$M = \frac{EK}{12(1 - \nu^2)} (T_{e(\text{crust})}^3 + T_{e(\text{mantle})}^3) = M_{\text{crust}} + M_{\text{mantle}}, \quad (2)$$

where E is the Young's modulus, K is the topography curvature, ν is the Poisson's coefficient, $T_{e(\text{crust})}$ and $T_{e(\text{mantle})}$ are the effective elastic thicknesses of the crust and mantle, respectively, and M_{crust} and M_{mantle} are the crust and mantle contributions to the total bending moment; the effective elastic thickness of the lithosphere is $T_e = (T_{e(\text{crust})}^3 + T_{e(\text{mantle})}^3)^{1/3}$.

We use diabase ([Caristan, 1982](#)) and dry olivine ([Chopra and Paterson, 1984](#)) rheologies for the crust and mantle, respectively. The elastic parameters are taken as $E = 100$ GPa and $\nu = 0.25$, for consistency with [McGovern et al. \(2002, 2004\)](#). (The calculations of effective elastic thicknesses are dependent on the used elastic parameters; the use of different values may change the obtained elastic thickness, but it does not necessarily changes the obtained heat flows in a substantial manner ([Ruiz, 2005](#)).) Topography curvature is taken as the maximum curvature of the elastic layer in [McGovern et al. \(2004\)](#): $6.4 \times 10^{-7} \text{ m}^{-1}$ for $T_e = 24$ km and $3.0 \times 10^{-7} \text{ m}^{-1}$ for $T_e = 37$ km. Curvatures are concave upward, and for that reason the brittle strength is calculated for compression according to the low-pressure Byerlee's rule for zero pore pressure and crust density of 2900 kg m^{-3} . Calculations are performed for strain rates of 10^{-19} and 10^{-16} s^{-1} , values bounding a range usually utilized in planetary research (e.g., [Solomon and Head, 1990](#); [McGovern et al., 2002](#)).

4. Heat flows from depth of faults

There is strong evidence of deep blind thrust faults beneath winkle ridges on several locations on Mars (e.g., [Schultz, 2000](#)). At Solis Planum, offsets between either sides of each winkle ridge are observed even after removing regional slope ([Golombek et al., 2001](#)). These offsets are maintained to the next ridge and are consistent with winkle ridge formation as shallow (likely backthrust) deformation above blind thrust faults, but are not consistent with exclusively thin skin models of winkle ridges formation. Since winkle ridge spacing at Solis Planum of ~ 50 km, and thrust faults having typically dip angles of 25° or higher (e.g., [Turcotte and Schubert, 2002](#)), the

faults should be at least 25 km deep beneath the surface at the next ridge ([Golombek et al., 2001](#)), which represents the minimum depth of faulting. This is consistent with winkle ridge origin by localization instability, because this mechanism predicts, for the formation time, a brittle–ductile transition (BDT) 30–50 km deep beneath the ridged highland plana ([Montesi and Zuber, 2003](#)).

The BDT indicates the depth at which brittle and ductile strength are equal. Due to the temperature dependence of ductile strength, the temperature at the BDT depth can be obtained, which can in turn be converted, for a given strain rate, to heat flows by matching to a temperature profile ([Ruiz and Tejero, 2000](#); [Nimmo and Watters, 2004](#); [Ruiz, 2005](#)), which is here constructed as indicated in Section 2. We use a minimum BDT depth of 25 km to establish upper limits for heat flow in the time of faulting, by calculating brittle (compressive) and ductile strengths in the crust as described in Section 2.

5. Results

Surface heat flows in terms of b , calculated for $T_e = 24$ –37 km and $A = 0.5$ – $0.7 \text{ } \mu\text{W m}^{-3}$, are shown as black curves in [Fig. 1](#). For a given value of surface heat flow, if b is thicker the heat flow contribution from the HPE-enriched crust is higher, and the mantle heat flow lower, which in turn implies colder temperatures at the lower lithosphere, increasing total lithospheric strength. For this reason, surface heat flow increases and mantle heat flow decreases with increasing b (i.e., with thickening HPE-enriched crust).

Surface heat flows in terms of b , calculated for BDT depth = 25 km and $A = 0.5$ – $0.7 \text{ } \mu\text{W m}^{-3}$, are shown as gray curves in [Fig. 1](#). Surface heat flow increases with b up to $b = \text{BDT depth}$, but it is constant for $b > \text{BDT depth}$. This occurs because the calculations exclusively depend on the temperature at, and distribution of heat sources above, the BDT depth. So, for $b > \text{BDT depth}$, increasing the product Ab implies lowering the lower crust and lithospheric mantle heat flow. For $A = 0.7 \text{ } \mu\text{W m}^{-3}$, the cases $b > 57$ km for 10^{-19} s^{-1} are unrealistic because they would imply negative mantle heat flow.

The two sets of independent heat flow calculations presented here are based on geological structures of roughly similar age, and should therefore be roughly consistent. As indicated above, heat flows obtained from a lower limit for the BDT depth are, by definition, upper limits. For this reason, for equal strain rate, HPE-enriched crust thickness b , and heat production rate A , elastic thickness-based values that are higher than BDT-based values should be rejected. This provides constraints on surface and mantle heat flows, along with upper (HPE-enriched) and lower (“HPE-poor”) crust thicknesses for Solis Planum in the Hesperian. It can be seen that for the majority of possible settings the permitted HPE-enriched crust is thinner than the whole crust. Indeed, a HPE-enriched crust as thick as the whole crust is only permitted for high effective elastic thickness and high strain rate, and low A . The results obtained are summarized in [Table 1](#) (left column).

[Fig. 2](#) shows the effect of varying A between 0.4 and $0.9 \text{ } \mu\text{W m}^{-3}$ in the calculations. This range is equivalent to

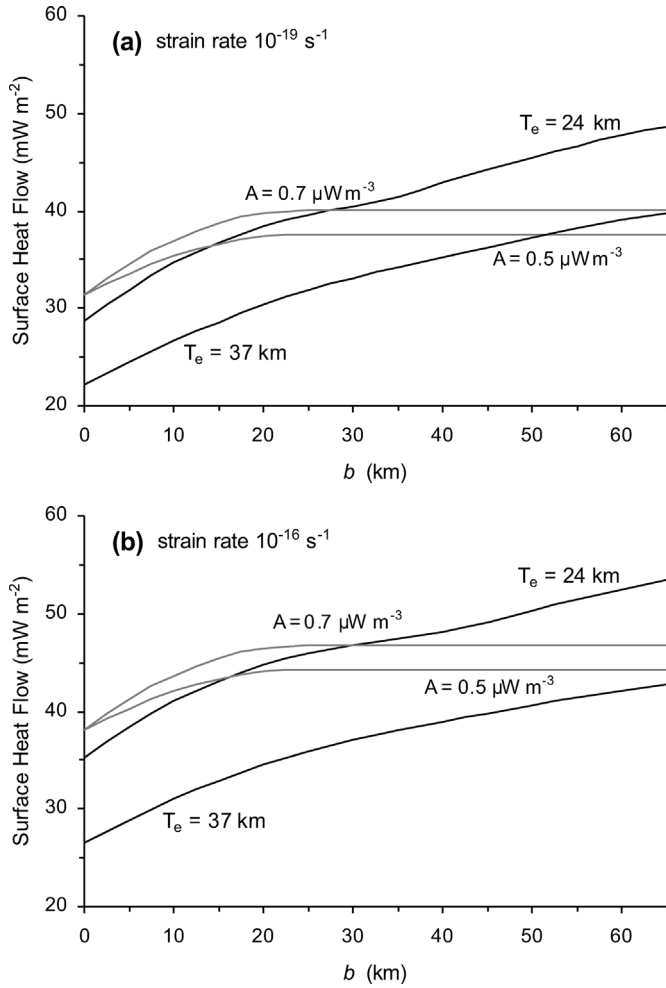


Fig. 1. Surface heat flow in terms of b , the thickness of the heat-producing elements (HPE)-enriched crust, calculated for strain rates of (a) 10^{-19} s^{-1} and (b) 10^{-16} s^{-1} . The black curves are obtained from the effective elastic thickness for the extreme cases of $T_e = 24 \text{ km}$ with $A = 0.7 \mu\text{W m}^{-3}$, and $T_e = 37 \text{ km}$ with $A = 0.5 \mu\text{W m}^{-3}$. Gray curves are upper limits calculated from a lower limit to the brittle-ductile transition (BDT) depth of 25 km. Labels for A are valid for both T_e - and BDT-based values.

Table 1
Summary of results for the nominal model

	Whole range of results	Results for $20 \leq b \leq 30 \text{ km}$
Surface heat flow (mW m^{-2})	22–47	30–47
Mantle heat flow (mW m^{-2})	12–38	18–31
Fraction of the surface heat flow due to crustal heat sources	≤ 0.49 –0.73	0.29–0.52
b (km)	≤ 28 –65	20–30
(Total crust thickness – b) (km)	≥ 0 –37	35–45

varying the K abundance between 0.2 and 0.7% (the approximate range deduced from Mars Odyssey GRS observations; Boynton et al., 2003; Taylor et al., 2003a, 2003b) while maintaining Th and U abundances as above, although it also serves to explore the effects of reasonable variations of A independently of the exact abundances of each element. This figure presents the b values for which the T_e -based heat flow is equal to the

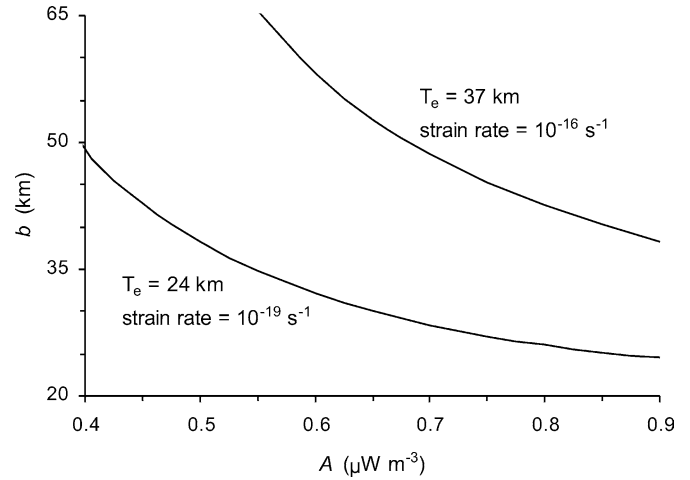


Fig. 2. HPE-enriched crust thickness (b) values for which the T_e -based heat flow is equal to the heat flow calculated from a BDT 25 km deep, shown in terms of A . Only are represented the extreme cases $T_e = 24 \text{ km}$ with strain rate $= 10^{-19} \text{ s}^{-1}$, and $T_e = 37 \text{ km}$ with strain rate $= 10^{-16} \text{ s}^{-1}$.

BDT depth-based heat flow. Only are represented the extreme cases $T_e = 24 \text{ km}$ with strain rate $= 10^{-19} \text{ s}^{-1}$, and $T_e = 37 \text{ km}$ with strain rate $= 10^{-16} \text{ s}^{-1}$, which cover the entire range of permitted b values for BDT depth = 25 km. It can be seen again that for most parameter combinations the HPE-enriched crust is thinner than the whole crust.

The arguments here presented are more pressing when it is recalled that 25 km is an upper limits for the BDT depth. A deeper BDT gives lower surface heat flows, and hence a thinner HPE-enriched crust. So, for BDT depth $> 26.1 \text{ km}$, it is obtained $b < 65 \text{ km}$ for any setting.

Two other effects may affect the results. The thermal conductivity of the crust could be somewhat lower than that used in this work, since basalts frequently have thermal conductivities close to $2 \text{ W m}^{-1} \text{ K}^{-1}$ (Beardsmore and Cull, 2001). In this case, both BDT- and T_e -based heat flows are decreased, but for the latest this reduction is minor in order to compensate the strength increasing due to a colder geotherm. The consequence would a thinner HPE-enriched crust. We have used for Solis Planum a crust thickness based in the crustal model in Neumann et al. (2004). A thicker (thinner) crust would reduces (increases) the strength at the mantle top, and consequently the required heat flow to justify a given effective elastic thickness is lower (higher), which in turn results in a higher (lower) permitted thickness for the HPE-enriched crust. Future modeling following this line of research should take into account these effects in more detail.

6. Implications

Our results strongly support a differentiated martian crust, since permitted HPE-enriched crustal thickness values are generally thinner than the whole crust at Solis Planum. This is consistent with the proposal of Norman (2002), based on the geochemistry of martian meteorites, that the LREE-enriched crust originated early in the history of Mars from an undepleted mantle, and that subsequent additions to the crust were de-

rived from depleted mantle sources. The LREE-enriched crust thickness (globally averaged) of ≤ 45 km is consistent with the HPE-enriched crust thickness obtained here, although it is worth emphasizing that this is a local value for Solis Planum.

If the most likely range of globally averaged thickness for the LREE-enriched crust, between 20 and 30 km, is considered roughly valid for Solis Planum, and if this range is also considered representative of the HPE-enriched crust, then a narrower constraint for thermal properties in the Hesperian could be tentatively proposed for this region (Table 1, right column). In this case, between 29 and 52% of the surface heat flow was due to crustal heat sources, an amount comparable to that in terrestrial continental regions.

In conclusion, geodynamic modeling of the lithosphere is capable of providing information about the differentiation state of the martian crust. We have found evidence for an upper HPE-enriched crust thinner than the whole crust in the Solis Planum region. This result is significant, and future refinement and extension of this kind of work will be very useful for understanding of the thermal and geochemical evolution of Mars.

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