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The present-day thermal state of Mars

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ABSTRACT

The present-day thermal state of the martian interior is a very important issue for understanding the internal evolution of the planet. Here, in order to obtain an improved upper limit for the heat flow at the north polar region, we use the lower limit of the effective elastic thickness of the lithosphere loaded by the north polar cap, crustal heat-producing elements (HPE) abundances based on martian geochemistry, and a temperature-dependent thermal conductivity for the upper mantle. We also perform similar calculations for the south polar region, although uncertainties in lithospheric flexure make the results less robust. Our results show that the present-day surface and sublithospheric heat flows cannot be higher than 19 and 12 mW m^{-2}, respectively, in the north polar region, and similar values might be representative of the south polar region (although with a somewhat higher surface heat flow due to the radioactive contribution from a thicker crust). These values, if representative of martian averages, do not necessarily imply sub-chondritic HPE bulk abundances for Mars (as previously suggested), since (1) chondritic composition models produce a present-day total heat power equivalent to an average surface heat flow of 14–22 mW m^{-2} and (2) some convective models obtain similar heat flows for the present time. Regions of low heat flow may even have existed during the last billions of years, in accordance with several surface heat flow estimates of ~20 mW m^{-2} or less for terrains loaded during Hesperian or Amazonian times. On the other hand, there are some evidences suggesting the current existence of regions of enhanced heat flow, and therefore average heat flows could be higher than those obtained for the north (and maybe the south) polar region.

1. Introduction

Knowledge of the surface heat flow is very important for understanding the thermal and geologic evolution of a planetary body. There are not direct measurements for Mars, but heat flows have been deduced for diverse martian regions from the effective elastic thickness of the lithosphere (Solomon and Head, 1990; Anderson and Grimm, 1998; Zuber et al., 2000; Nimmo, 2002; Kiefer, 2004; McGovern et al., 2002, 2004; Grott et al., 2005; Ruiz et al., 2006, 2008; Kronberg et al., 2007; Ruiz, 2009; Dohm et al., 2009a; Ritzer and Hauck, 2009) or from the depth to the brittle–ductile transition beneath large thrust faults (Schultz and Watters, 2001; Grott et al., 2007; Ruiz et al., 2008, 2009). So deduced heat flows are valid for the time when the lithosphere was loaded or faulted, permitting us to delineate, in a first approximation, the thermal evolution of the planet (McGovern et al., 2002, 2004; Montesi and Zuber, 2003).

Given the lack of large scale tectonic activity at the present time, the possibility of using effective elastic thicknesses for estimating the present-day thermal state of this planet seems restricted to the polar regions, where loading by ice caps is a recent phenomenon, estimated to be a few million years old (Laskar et al., 2002; Phillips et al., 2008). Through the modeling of the deflection of the topography beneath the north polar cap due to ice loading, Phillips et al. (2008) found a lower limit of 300 km for the effective elastic thickness of the lithosphere at the north polar region. For the south polar region, Phillips et al. (2008) suggested a lower limit of 275–300 km for the effective elastic thickness, although the topographic evidences for this case are more uncertain. Effective elastic thicknesses around or higher than 300 km are clearly higher than any previously published value for Mars (e.g., McGovern et al., 2004). This could indicate that Mars is dead in a geodynamical sense (Grott, 2008).

On the other hand, Wieczorek (2008) obtained a best fit of 161 km for the effective elastic thickness of the south polar region through gravity/topography admittance modeling. However, this

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author found that any value higher than 110 km can fit the observed admittance. A value of ~100 km or higher is similar to the effective elastic thicknesses obtained for the Hesperian/Amazonian-aged Valles Marineris region (McGovern et al., 2004), although the estimate for the south polar region is a lower limit. On the other hand, effective elastic thickness estimates for Amazonian-aged regions are usually lower than 100 km (McGovern et al., 2004; Belleguic et al., 2005), although these estimates were performed for volcanic regions, which could be hotter than average.

Phillips et al. (2008) estimated heat flows following the strength envelope procedure (McNutt, 1984) coupled with temperature profiles and crustal heat production rates deduced from the thermal history models of Hauck and Phillips (2002), obtaining surface and mantle heat flows of 25 and 17 mW m^{-2}, respectively. These results are very similar to those obtained by the nominal model of Hauck and Phillips (2002), and Phillips et al. (2008) conclude that, since the result for the north polar region is an upper limit, heat-producing elements (HPE) abundances for Mars are probably sub-chondritic (this conclusion was basically unchanged when a non-steady-state loading response was considered by these authors). In this sense, Grott and Breuer (2009a) found that thermal history models consistent with an average effective elastic thickness of at least 300 km require either sub-chondritic HPE abundances or a high fractionation of HPE in the crust and a dry mantle rheology. An alternative explanation is a spatially heterogeneous heat flow on Mars, which could naturally arise from mantle convection (Grott and Breuer, 2009a; Kiefer and Li, 2009).

The conclusions of Phillips et al. (2008) rely on HPE distributions and temperature profiles derived from thermal history models, and somewhat on the assumed mantle thermal conductivity (4 W m^{-1} K^{-1}), which is high for the lithospheric mantle. Here we derive a more robust calculation of upper limits for the surface and sublithospheric heat flow at both polar regions by using a HPE distribution based on geochemical considerations and more realistic thermal conductivities, including a temperature-dependent thermal conductivity for the lithospheric mantle; this procedure has the advantage of making the results independent of any specific thermal history model. We also discuss the implications of our results for the thermal history of Mars, as well as some possible evidences of regional variations in the current thermal state of the lithosphere.

2. Temperature at the base of the lithosphere

The effective elastic thickness of the lithosphere, which is a measure of the total strength of the lithosphere, can be converted to heat flow following the equivalent strength envelope procedure described by McNutt (1984). This methodology is based on the condition that the bending moment of the mechanical lithosphere must be equal to the bending moment of the equivalent elastic layer of thickness T_e. The link between the mechanical structure and heat flow comes from the dependence of the ductile strength on temperature.

If lithospheric flexure is small, then it can be neglected and T_e is equal to the depth to the base of the mechanical lithosphere, L, which is defined as the depth at which the ductile strength reaches a low value and below which there are no further significant increases in strength. The base of the mechanical lithosphere can be defined, therefore, as the depth to an isotherm given by

$$T_L = \frac{Q}{R \ln\left[\frac{A(\sigma_1-\sigma_3)_L^n}{\dot{\varepsilon}}\right]}, \tag{1}$$

where Q is the activation energy for creep, R (=8.31447 J mol^{-1} K^{-1}) is the gas constant, A and n are laboratory-determined constants, $(\sigma_1-\sigma_3)_L$ is the strength level defining the base of the mechanical lithosphere, and $\dot{\varepsilon}$ is the strain rate. The surface, subcrustal and sublithospheric heat flows can then be calculated by finding the temperature profile that fits the isotherm given by Eq. (1) at a depth equal to T_e. Surface heat flow decreases with topography curvature (McNutt, 1984; Solomon and Head, 1990), and therefore assuming no lithosphere flexure gives an upper limit to the heat flow. This is sufficient for the purposes of this work, as Phillips et al. (2008) did not determine a lower limit for lithospheric flexure.

We calculate the temperature defining the base of the mechanical lithosphere by using the flow law of dry olivine (Chopra and Paterson, 1984): Q = 535 kJ mol^{-1}, A = 28840 MPa^{-n} s^{-1}, and n = 3.6. For $(\sigma_1-\sigma_3)_L$ we use a value of 10 MPa, which is considered appropriate for the low gravity of Mars (see Ruiz et al., 2006), although the exact selected value does not produce significant changes in the calculations due to the exponential dependence of ductile strength on temperature. A strain rate of 10^{-14} s^{-1} is used in the calculations, which corresponds to an estimated age of ~5 Ma for the load by the north polar cap (Phillips et al., 2008). So, T_L = 1267 K is obtained. The use of a weaker wet olivine rheology would reduce both the temperature at the base of the lithosphere and the surface heat flow. For example, for the wet Anita Bay dunite (Chopra and Paterson, 1984) T_L = 1087 K, which would imply a reduction of less than 10% of the surface heat flow obtained in all the cases analyzed in the present work (see Sections 4 and 5).

3. Temperature profiles

We calculate thermal profiles for both north and south polar regions by considering a three layer model differencing between polar cap (not included in the lithosphere), crust and mantle lithosphere.

Polar caps are assumed to be composed of water–ice. The thermal conductivity of cold water–ice is high, and the presence of rocks or other ices (e.g., CO_2) would reduce the bulk thermal conductivity, and hence the calculated heat flow. For this reason, non-water–ice components are not taken into account in our upper-limit calculation. The thermal conductivity of water–ice is strongly temperature-dependent, and therefore the temperature profile in the polar cap is given by

$$T_{pcb} = T_s \exp\left(\frac{F b_{pc}}{k_0}\right), \tag{2}$$

where T_s is the surface temperature, F is the surface heat flow (equal to the heat flow reaching the polar cap from below), b_{pc} is the thickness of the polar cap, and k_0 = 621 W m^{-1} (Petrenko and Whitworth, 1999). Here we use T_s = 155 K and b_{pc} = 2 km as representatives for the martian polar regions (Plaut et al., 2007; Phillips et al., 2008; Wieczorek, 2008). Subsurface temperatures might be higher than those expected for a steady-state thermal profile, because cooler surface conditions related to obliquity and insolation changes implied in the growth of the polar caps (Laskar et al., 2002) would have a relatively limited downward propagation in a few million years; this would reduce the surface heat flow, reinforcing the robustness of the upper limits calculated in this work.

For the crust, homogeneously distributed radioactive heat sources and a constant thermal conductivity are assumed. So, the temperature at the base of the crust is given by

$$T_{cb} = T_{pcb} + \frac{F b_c}{k_c} - \frac{H_c b_c^2}{2k_c}, \tag{3}$$

where T_{pcb} is the surface temperature at the base of the polar cap, b_c is the thickness of the crust, k_c is the thermal conductivity of the crust, and H_c is the crustal volumetric heat production rate.

Crustal models for Mars (Neumann et al., 2004, 2008) indicate that the crust at the north polar region is typically ~15–25 km