Interior Dynamics and Thermal Evolution of Mars – a Geodynamic Perspective

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Abstract

Over the past decades, global geodynamic models have been used to investigate the thermal evolution of terrestrial planets. With the increase of computational power and improvement of numerical techniques, these models have become more complex, and simulations are now able to use a high resolution 3D spherical shell geometry and to account for strongly varying viscosity, as appropriate for mantle materials. In this study we review global 3D geodynamic models that have been used to study the thermal evolution and interior dynamics of Mars. We discuss how these models can be combined with local and global observations to constrain the planet’s thermal history. In particular, we use the recent InSight estimates of the crustal thickness, upper mantle structure, and core size to show how these constraints can be combined with 3D geodynamic models to improve our understanding of the interior dynamics, present-day thermal state and temperature variations in the interior of Mars. Our results show that the crustal thickness variations control the surface heat flow and the elastic thickness pattern, as well as the location of melting zones in the present-day martian mantle. The lithospheric temperature and the seismic velocities pattern in the shallow mantle reflect the crustal thickness pattern. The large size of the martian core leads to a smaller scale convection pattern in the mantle than previously suggested. Strong mantle plumes that produce melt up to recent times become focused in Tharsis and Elysium, while weaker plumes are distributed throughout the mantle. The thickness of the seismogenic layer, where seismic events can occur, can be used to discriminate between geodynamic models, if the source depth and location of seismic events is known. Furthermore model predictions of present-day martian seismicity can be compared to the values measured by InSight. Future models need to consider recent estimates from the present-day elastic lithosphere thickness at the north pole of Mars, the effects of lateral variations of seismic velocities on waves propagation through the mantle and lithosphere, and to test the spatial distribution of seismicity by comparing model predictions to observations.

Keywords: Geodynamic modeling, Mars, Thermal evolution, InSight, Surface heat flow, Elastic thickness, Tidal deformation, Seismic velocities, Seismogenic layer

1 Introduction

A variety of numerical modeling studies have been used to investigate the thermal evolution and interior dynamics of Mars (see Breuer and Moore, 2015; Smrekar et al., 2019 for a review). These studies use either fully dynamical 2D/3D simulations that self-consistently model the evolution of mantle flow and through their nature can address local output quantities in regions of interest or 1D parametrized models that track the evolution of global quantities, such as the global mantle temperature or the average surface heat flow. Fully dynamical 2D and 3D models have been used to predict the surface heat flow and seismicity distribution on present-day Mars (Plesa et al., 2016; 2018a), the effects of impacts during the early thermochemical history (Ruedas and Breuer, 2017; Roberts and Arkani-Hamed, 2017), partial melting and crust-mantle differentiation (Ruedas et al., 2013; Plesa and Breuer, 2014), and the formation of the martian crustal thickness dichotomy (Keller and Tackley, 2009; Golabek et al., 2011). On the other hand, 1D parametrized thermal evolution models have been applied to investigate the crustal formation and crust-mantle differentiation (Hauck and Phillips, 2002; Breuer and Spohn, 2006; Morschhauser et al., 2011), as well as mantle degassing (Fraeman and Korenaga, 2010; Grott et al., 2011), the magnetic field evolution (Breuer and Spohn, 2000; Williams and Nimmo, 2004), the thermal state of the lithosphere (Grott and Breuer, 2008; Thiriet et al., 2018), and the coupled thermal-orbital evolution of Mars and its moon Phobos (Samuel et al., 2019).
With the arrival of the InSight mission in 2018 (Banerdt et al., 2020), a new chapter of planetary geophysics has begun. Equipped with a seismometer (SEIS, Lognonné et al., 2020), a heat flow probe (HP^3, Spohn et al., 2018) and X-band telecommunication capabilities that allow to precisely measure the rotation of Mars (RISE, Folkner et al., 2018), InSight investigates the interior of Mars from its core to the surface.

InSight’s measurements represent the most direct set of constraints for the interior of Mars. The crustal thickness (Knapmeyer-Endrun et al., 2021), the thickness and thermal state of the lithosphere (Khan et al., 2021), and the size of the core (Stähler et al., 2021) provide important information for modeling of thermal evolution and investigating physical processes active in the interior of Mars. In turn, numerical models of the thermal evolution combined with InSight measurements can be used to constrain poorly known parameters, such as the rheology of the mantle, distribution of heat producing elements (HPEs) between the mantle and crust, and the evolution of the surface and core-mantle boundary (CMB) heat flows. Previous 1D models that used inversion techniques for the interior structure, composition, and thermal state of the mantle or made assumptions of the present-day thermal structure have made predictions about the seismic structure of the martian mantle (e.g., Khan et al., 2018; Zheng et al., 2015; Gudkova and Zharkov, 2004; Sohl and Spohn, 1997; Mocquet et al., 1996). More recent thermal evolution models have directly addressed the seismic observations of InSight. These models investigated the effects of crustal thickness and its enrichment in HPEs on the thermal evolution and present-day partial melt distribution in the interior of Mars (Knapmeyer-Endrun et al., 2021), studied the consequences of a molten layer at the base of the mantle on the thermal history and core size measurements (Samuel et al., 2021), and estimated seismic velocities variations due to the interior temperature distribution (Plesa et al., 2021).

In this study we focus on 3D thermal evolution models and discuss results presented in previous studies and new simulations in the framework of the recent InSight data about the martian crust, mantle, and core. In Section 2 we review the mathematical equations used by geodynamic models to calculate the thermal evolution of Mars. Section 3 presents the effects of the crustal thickness variations on model predictions for heat flow and elastic lithospheric thickness variations, as well as on the distribution of partial melt zones in the mantle. The dynamics inside the mantle, mantle seismic velocities variations due to mantle thermal anomalies, and constraints from tidal deformation are discussed in Section 4. In Section 5 we list the core size estimates from geodetic data and seismic observations and discuss their consequence on the mantle dynamics and convection pattern. The seismogenic layer thickness and the present-day seismicity are discussed in Section 6. In the last section (Section 7) we present a summary of geodynamic models and their findings and suggest future investigations.

We discuss in detail selected models that are compatible with the crustal thickness and core radius estimates that were derived from InSight’s seismic measurements. Throughout this study we illustrate how model results can be related to available geological and geophysical observations. For our models we use both global constraints such as the thermal lithosphere thickness (Khan et al., 2021) and estimates of the tidal deformation of Mars (Genova et al., 2016; Konopliv et al., 2016, 2020), as well as local data sets, such as the present-day elastic lithospheric thickness at the north and south poles (Phillips et al., 2008; Wieczorek, 2008; Broquet et al., 2020, 2021) and the location of partial melt that could explain recent volcanic activity in Tharsis and Elysium volcanic centers. We highlight how our knowledge of the thermal history and present-day state of the interior has improved with the recent InSight results and discuss what future models need to address in order to further constrain the thermal evolution of Mars. Current geodynamic modeling will also help to identify scientifically interesting landing sites for a future seismic network that may become available after 2030 (Stähler and Knapmeyer, 2022).

2 Geodynamic modeling and thermal evolution of Mars

Geodynamic models that are used to investigate the interior dynamics of rocky planets in general, and of Mars in particular, numerically solve for a set of conservation equations. While a 3D spherical geometry is the choice for investigating the convection pattern in the interior and relating the results to local observations, 2D models have the advantage of being computationally faster making it possible to test a larger parameter space with a higher spatial resolution compared to their 3D counterparts. The equations and parameters used in these models are typically nondimensional. They are scaled with the mantle thickness \(D\) as length scale, a reference thermal diffusivity \(\kappa\) as time scale, and the initial temperature difference \(\Delta T\) as temperature scale, although some codes use dimensional quantities (Kronbichler et al., 2012; Heister et al., 2017). The advantage of the nondimensionalization is that characteristic dimensionless numbers such as the Rayleigh number, which is defined as the ratio between parameters driving convection and those opposing it, can be used to describe the convection system without needing to know the exact detailed parameters of a simulation.

The system of equations that is solved includes the conservation of mass, linear momentum and thermal energy (e.g., Schubert et al., 2001). Their nondimensional formulation for a system assuming a Newtonian rheology, an
infinite Prandtl number as appropriate for high viscosity media with negligible inertia, a variable thermal expansivity and conductivity, using the Extended Boussinesq Approximation (EBA), and including solid-solid phase transitions read (e.g., Christensen and Yuen 1985):

\[ \nabla \cdot \mathbf{u} = 0, \quad (1) \]

\[ \nabla \cdot [\eta(\nabla \mathbf{u} + (\nabla \mathbf{u})^T)] - \nabla p + (Ra \alpha T - \sum_{l=1}^{3} R_b \Gamma_l) \varepsilon_r = 0, \quad (2) \]

\[ \frac{dT}{Dt} - \nabla \cdot (k \nabla T) - Di \alpha (T + T_0) u_r - \frac{Di}{H} \Phi - \sum_{l=1}^{3} Di R_b \frac{D \Gamma_l}{Dt} \gamma_l (T + T_0) - H = 0, \quad (3) \]

where \( \mathbf{u} \) is the velocity vector, \( u_r \) is its radial component, \( \eta \) is the viscosity, \( p \) is the dynamic pressure, \( \alpha \) is the thermal expansivity, \( T \) is the temperature, \( \varepsilon_r \) is the unit vector in radial direction, \( t \) is the time, \( k \) is the thermal conductivity, \( Di \) is the dissipation number, and \( \Phi \equiv \tau : \dot{\varepsilon} / 2 \) is the viscous dissipation, where \( \tau \) and \( \dot{\varepsilon} \) are the deviatoric stress and strain-rate tensors, respectively.

The Rayleigh number that describes the vigor of convection, the internal heating rate \( H \) that controls the mantle heating due to radioactive elements (HPEs), and the dissipation number \( Di \) that accounts for the increase of temperature due to adiabatic compression effects are defined as follows:

\[ Ra = \frac{\rho g \Delta T D^3}{\eta \kappa}, \quad H = \frac{\rho Q_{HPE} D^2}{k \Delta T}, \quad Di = \frac{\alpha g D}{c_p}, \quad (4) \]

where \( g \) is the gravitational acceleration, \( c_p \) is the mantle heat capacity, and \( Q_{HPE} \) is the heat production rate in W kg\(^{-1}\).

Previous geodynamic models accounted for two exothermic phase transitions, and, in case of a small core radius, for an additional endothermic phase change. While exothermic phase transitions tend to accelerate mantle flow, the endothermic one slows it down and promotes layered convection. In particular, the endothermic phase transition from wadsleyite/ringwoodite to bridgmanite has been suggested to affect the style of convection. This phase transition significantly changes the CMB heat flow and leads to a low degree convection pattern that was proposed by previous studies to explain the formation of the Tharsis volcanic province (Harder and Christensen 1996; Breuer et al. 1998; Spohn et al. 1998; Van Thienen et al. 2006). However, this phase transition is not relevant for models with a core radius of 1700 km or larger, and in these scenarios the convection in the mantle is characterized by more than one plume (Spohn et al. 1998; Michel and Forni 2011).

One of the most important parameters in geodynamic models is the mantle viscosity, which is temperature and pressure dependent and follows an Arrhenius law. Its nondimensional formulation (e.g., Roberts and Zhong 2006) reads:

\[ \eta(T, z) = \exp \left( \frac{E + z V}{T + T_0} - \frac{E + z_{ref} V}{T_{ref} + T_0} \right), \quad (5) \]

where \( z \) is the depth, \( E \) and \( V \) are the activation energy and activation volume, respectively, \( T_0 \) is the nondimensional surface temperature, and \( z_{ref} \) and \( T_{ref} \) are the reference depth and temperature where the reference viscosity is attained. The temperature dependence of the viscosity is controlled by the activation energy while the pressure dependence by the activation volume (e.g., Karato and Wu 1993; Hirth and Kohlstedt 2003).

At the temperature and pressure conditions of planetary mantles the viscosity varies over orders of magnitude and is the parameter primarily controlling the vigor of convection, the formation of a stagnant lid – an immobile layer at the top of the convecting mantle caused by the strong increase of viscosity with decreasing temperature –, and the convection pattern in the mantle. The latter is sensitive to the increase of the viscosity with depth, and may result in a low degree convection pattern for a strong depth-dependent viscosity or a sudden viscosity increase (i.e., a viscosity jump) in the mid-mantle (Roberts and Zhong 2006; Keller and Tackley 2009). Such a convection pattern has been previously proposed to explain the crustal thickness dichotomy and focused volcanic activity in Tharsis, which is the largest volcano-tectonic region on Mars (e.g., Harder and Christensen 1996; Breuer et al. 1998; Zhong and Zuber 2001; Roberts and Zhong 2006; Keller and Tackley 2009).

Geodynamic thermal evolution models account for the decay of radioactive elements with time and employ a cooling boundary condition at the CMB. Using a 1-D energy balance the evolution of the CMB temperature is calculated under the assumption of a constant core density and heat capacity (Stevenson et al. 1983; Steinbach and Yuen 1994):

\[ c_c \rho_c V_c \frac{dT_{CMB}}{dt} = -q_c A_c, \quad (6) \]

where \( c_c \), \( \rho_c \), and \( V_c \) are the core heat capacity, core density, and core volume, respectively. \( T_{CMB} \) and \( q_c \) are the temperature and the heat flow at the CMB, respectively, while \( A_c \) is the CMB area.
Geodynamic models have been employed in previous studies to investigate the effects of crustal thickness variations, as derived from gravity and topography data, on the surface heat flow variations and the thermal state of the lithosphere (Plesa et al., 2016, 2018b). The crust in these studies does not change with time, but varies spatially according to the chosen crustal thickness model (Wieczorek and Zuber, 2004, Wieczorek et al., 2022). Results show that the crustal thickness variations and the crustal enrichment in heat producing elements (HPEs) control the surface heat flow distribution at present day, the thickness of the lithosphere and the lithospheric temperature variations. In addition to global output quantities such as the average lithosphere thickness, average surface and CMB heat flows and average mantle temperature, these models can provide local values that can be compared to regional estimates. This is essential to evaluate constraints provided by local measurements and helps to put regional scale data in a global context.

3 Crustal thickness estimates, partial melting, and the thermal state of the lithosphere

The thickness of the crust provides important constraints for later planetary differentiation after core formation and for the overall thermal evolution of the mantle. The crust that is built after the initial differentiation of the planet and the crystallization of a potential magma ocean records the magmatic activity through time, as it is formed by partial melting of the mantle. During mantle melting, incompatible elements such as heat producing elements (U\(^{235,238}\), K\(^{40}\), and Th\(^{232}\)) and volatiles (e.g., \(\text{H}_2\text{O}\) and \(\text{CO}_2\)) are preferentially enriched in the melt. The melt, due to its lower density compared to the surrounding mantle, rises to the surface, where it crystallizes and produces the crust. While volatiles, such as \(\text{H}_2\text{O}\) and \(\text{CO}_2\), are released in the atmosphere, the heat producing elements (HPEs) remain stored in the crust that becomes more enriched than the primitive mantle.

On Mars, the bulk of the crust has been built during the early history (Greeley and Schneid, 1991, Nimmo and Tanaka, 2005) with an intense volcanic activity during Noachian. Over time, volcanic activity declined and became more focused in Tharsis and Elysium, the largest volcanic provinces on Mars. Young lava flows in both Tharsis (Neukum et al., 2004, Hauber et al., 2011) and Elysium (Vaucher et al., 2009) indicate that Mars has remained volcanically active over most of its history and that partial melt production in the planet’s interior may be ongoing. The crustal heat production rate that was derived from the surface abundances of thorium and potassium recorded by the gamma-ray spectrometer (GRS) on board Mars Odyssey (Taylor et al., 2006a) indicates a higher crustal enrichment factor compared to typical values for mid-ocean ridge basalts on Earth (Kaula et al., 1981). The GRS data shows also a rather homogeneous distribution with crustal thorium abundances between 0.2 and 1 ppm (Taylor et al., 2006b), indicating much smaller variations than those observed on the Moon (Lawrence et al., 2000). Based on this rather homogeneous distribution of thorium and potassium at the surface of Mars and assuming that the surface abundance of HPEs is representative for the entire crust, it was suggested that crustal thickness variations have a stronger effect on the crustal heat flow variations (Hahn et al., 2011).

Perhaps the most prominent geological feature on Mars is the crustal dichotomy. The cause for the difference in elevation and crustal thickness between the southern highlands and the northern lowlands is poorly known. Crustal thickness models that can explain the gravity and topography data are nonunique. While an anchor point given by a crustal thickness value at a known location can help to constrain these models, one major assumption that remains is the density of the crust and how it varies laterally (Wieczorek et al., 2022). When considering different crustal densities for the northern lowlands and southern highlands, the difference in crustal thickness across the dichotomy boundary can be small (Fig. 1a), entirely absent (Fig. 1b), or clearly visible (Fig. 1c).

Seismic data from the InSight mission have been used to determine the crustal thickness at the InSight landing site (Knapmeyer-Endrun et al., 2021). In the initial receiver function analyses, which are sensitive to the local crustal structure beneath the Insight lander (Knapmeyer-Endrun et al., 2021), two possible crustal models were premissible: a two layer crust based on two strong seismic discontinuities below the surface at depths of about 8 and 20 km, or a three layer crust that included a third weaker discontinuity recorded in the data at about 39 km depth. These seismic constraints can be used in combination with gravity and topography data to construct global crustal thickness models of the planet (Wieczorek and Zuber, 2004, Wieczorek et al., 2022). For the two layer model, such modeling predicts an average crustal thickness somewhere between 24 and 38 km (Knapmeyer-Endrun et al., 2021), whereas for the three layer model the average thickness is predicted to be between 30 and 72 km (Wieczorek et al., 2022). We note that the range of crustal thicknesses in Wieczorek et al., 2022 for the three layer model is slightly larger than the range of 39 to 72 km presented in Knapmeyer-Endrun et al., 2021. This is because the initial study considered that the density of the crust was laterally homogeneous, whereas the latter study considered cases where the density of the crust could differ across the dichotomy boundary.

While none of the two crustal thickness models (two layer or three layer) can be currently excluded based on
Figure 1: Distribution of the crustal thickness (a, b, and c), associated heat flow variations (d, e, and f) and elastic lithosphere thickness (g, h, and i) at present day. Panels a, d, and g show the thin crustal thickness end-member with an average crustal thickness of 40.6 km and a crustal density of 2550 kg m\(^{-3}\). Panels b, e, and h present a crustal thickness model with an average thickness of 43.1 km and a density difference between northern lowlands (3000 kg m\(^{-3}\)) and southern highlands (2600 kg m\(^{-3}\)). In panels c, f, and i an end-member model with an average crustal thickness of 71.4 km and a crustal density of 3000 kg m\(^{-3}\). The white and gray contour lines show the 0 km level of surface topography obtained from the Mars Orbiter Laser Altimeter (MOLA) on board Mars Global Surveyor (MGS).

In the receiver function analysis, thermal evolution models typically produce a thicker crust than the average crustal thickness predicted by the two layer model (Knapmeyer-Endrun et al., 2021). Moreover, a recent study by Kim et al. (2021) that analyzed free-surface multiples of the P-wave and combined these with receiver function analysis also favors the three layer crust scenario. Therefore here we will discuss only models using the three layer crust scenario.

In Fig. 1 we show three crustal thickness models that match the gravity and topography data, and that are anchored at the InSight landing site by the three layer crustal thickness derived from the seismic observations. The crustal thickness models represent end-members in terms of crustal densities and crustal thicknesses. The thinnest crust uses a crustal thickness at the InSight landing site of 31 km and the lowest considered crustal density of 2550 kg m\(^{-3}\) (Fig. 1a) leading to an average crustal thickness of 40.6 km. An end-member model with a crustal thickness of 49 km at the InSight landing site and the highest crustal density value of 3000 kg m\(^{-3}\) (Fig. 1c) leads to an average crustal thickness of 71.4 km. In addition to the two end-member models, Fig. 1b shows another model that uses a crustal thickness at InSight landing site of 47 km and different densities for the northern lowlands (3000 kg m\(^{-3}\)) and southern highlands (2600 kg m\(^{-3}\)). This model has an average crustal thickness of 43.1 km. In this case, the dichotomy in crustal density largely erases the crustal thickness dichotomy that is prominent in the other two models in Fig. 1a and c.

Previously, crustal thickness models have been combined with geodynamic models to investigate the effects of crustal thickness variations on the surface heat flow variations and thermal state of the lithosphere (Plesa et al., 2016, 2018b). These models show that crustal thickness variations are the main contributions to the heat flow variations and mantle plumes have only a minor effect (Plesa et al., 2016). Here we discuss 3D thermal evolution calculations...
similar to Plesa et al. (2016, 2018b) that include updated crustal thickness models, which have been constrained by InSight seismic data. A list of the parameters used in the geodynamic simulations is shown in Table 1 while details of the crustal thickness modeling are discussed in Wieczorek et al. (2022) and crustal parameters are shown in Table 2.

Table 1: Parameters used in the geodynamic models in Fig. 1

| Symbol | Description       | Value             |
|--------|-------------------|-------------------|
| $D$    | Mantle thickness  | 1550 km           |
| $T_{ref}$ | Reference temperature | 1600 K         |
| $p_{ref}$ | Reference pressure | $3 \times 10^9$ Pa |
| $E$    | Activation energy | $3 \times 10^5$ J mol$^{-1}$ |
| $V$    | Activation volume | $10 \times 10^{-6}$ m$^3$ mol$^{-1}$ |
| $T_{init}$ | Initial mantle temperature | 1800 K         |
| $\Delta T$ | Initial temperature drop across the mantle | 2000 K         |
| $\alpha$ | Reference thermal expansivity | $2.5 \times 10^{-5}$ K$^{-1}$ |
| $\eta$ | Reference viscosity | $10^{21}$ Pa s   |
| $c_p$  | Mantle heat capacity | 1142 J kg$^{-1}$ K$^{-1}$ |
| $\rho_c$ | Core density | 6000 kg m$^{-3}$ |
| $\rho_c$ | Core heat capacity | 850 J kg$^{-1}$ K$^{-1}$ |
| $g$    | Surface gravity acceleration | 3.72 m s$^{-2}$ |
| $k$    | Mantle thermal conductivity | 4 W m$^{-1}$ K$^{-1}$ |
| $k_{cr}$ | Crust thermal conductivity | 3 W m$^{-1}$ K$^{-1}$ |
| $\kappa$ | Mantle thermal diffusivity | $1 \times 10^{-6}$ m$^2$ s$^{-1}$ |
| $Q$    | Total initial radiogenic heating (mantle and crust) | $23.33 \times 10^{-12}$ W kg$^{-1}$ |

Table 2: Parameters of the crustal thickness models shown in Fig. 1. A detailed description of the crustal thickness models is presented in Wieczorek et al. (2022). $\rho_N$ and $\rho_S$ are the densities of the northern lowlands and southern highlands, respectively, $d_c$ is the average crustal thickness, min. $d_c$ and max. $d_c$ are the minimum and the maximum crustal thickness values, and $d_{c,InSight}$ is the crustal thickness at InSight location.

| Model                | $\rho_N$ [kg m$^{-3}$] | $\rho_S$ [kg m$^{-3}$] | avg. $d_c$ [km] | min. $d_c$ [km] | max. $d_c$ [km] | $d_{c,InSight}$ [km] |
|----------------------|------------------------|------------------------|-----------------|-----------------|-----------------|----------------------|
| thin crust           | 2550                   | 2550                   | 40.6            | 11.7            | 72.4            | 31                   |
| density dichotomy crust | 3000                 | 2600                   | 43.1            | 2.9             | 130.4           | 47                   |
| thick crust          | 3000                   | 3000                   | 71.4            | 5.0             | 157.2           | 49                   |

In the absence of direct heat flow measurements, the elastic lithosphere thickness at various times and locations can be used as a proxy for the surface heat flow, as it can be linked to the thermal state of the lithosphere. The elastic thickness characterizes the stiffness of the lithosphere in response to loading and can be related to the mechanical thickness, given a rheological model. The mechanical thickness is directly linked to the thermal state of the lithosphere, since it can be identified with an isotherm (McNutt, 1984) and thus, it can be directly compared to lithospheric temperatures from thermal evolution models (Fig. 1). This comparison can be performed at various times during the evolution and at different locations, depending on the time and location, for which elastic thickness estimates are available.

The base of the mechanical lithosphere can be calculated following the approach of Grott and Breuer (2008) and Plesa et al. (2016, 2018a) and assuming a bounding stress of $\sigma_B$ of the order of $10^7$ Pa (Grott and Breuer, 2010; Burov and Diament, 1995), which gives the temperature associated with ductile failure:

$$T_e = \frac{E}{R} \left[ \log \left( \frac{\sigma_B^n A}{\dot{\varepsilon}} \right) \right]^{-1},$$

where $E$, $A$, and $n$ are rheological parameters, $R$ is the gas constant, and $\dot{\varepsilon}$ is the strain rate. A list of the rheological parameters used to calculate the mechanical thickness is shown in Table 3.
The mechanical thickness represents an upper bound for the elastic lithosphere thickness. However, it should be noted that the mechanical and elastic thickness are similar for small curvatures and bending moments as it is the case for the large geological features that are considered here (McGovern et al., 2004; Belleguic et al., 2005). Thus in the following, we will use the term “elastic thickness”.

The elastic thickness of the mantle and crust can be determined using Eq. (7) for their individual rheological parameters (Table 3). For the calculations presented here, we use parameters for a dry olivine mantle and a wet diabase crust similar to Grott and Breuer (2008). These rheological parameters have been found to match best the elastic thickness estimates available for the early history (Noachian epoch) and present-day Mars (Grott and Breuer, 2010; Breuer et al., 2016; Plesa et al., 2018b). If a layer of incompetent crust separates the elastic cores of the mantle $D_m$ and crust $D_c$, then the elastic thickness of the crust and mantle system is significantly reduced and the effective elastic thickness can be calculated as follows (Grott and Breuer, 2008, 2010):

$$D_e = (D_m^3 + D_c^3)^{\frac{1}{3}}.$$  

(8)

Otherwise, if the elastic thickness of the crust equals the crustal thickness, then the effective elastic thickness is the sum of the two contributions.

On Mars, gravity and topography analysis, lithospheric flexure studies, and estimates of the brittle to ductile transition indicate elastic lithosphere thicknesses smaller than about 25 km during the Noachian epoch (Grott et al., 2013; Thiriet et al., 2018; Plesa et al., 2018b). On the other hand, present-day elastic thickness estimates that are available for the north and south poles of Mars indicate a much thicker and colder lithosphere at these two locations. This has been concluded based on the lack of downward deflection with uncertainties of 100–200 m, beneath the north polar cap as seen by the MARSIS and SHARAD radars (Phillips et al., 2008), while for the south pole a maximum lithospheric flexure of 770 m has been found (Broquet et al., 2021). Previous elastic lithosphere thickness estimates with values larger than 300 km for the north pole (Phillips et al., 2008) and larger than 150 km for the south pole (Wieczorek, 2008) have been reevaluated in two recent studies (Broquet et al., 2020, 2021). The latest estimates indicate an elastic thickness between 330 km and 450 km for the north pole of Mars (Broquet et al., 2020) and a value larger than 150 km with a best fit of 360 km for the south pole (Broquet et al., 2021).

The present-day elastic thickness estimates at the north and south poles of Mars represent some of the strongest constraints for the thermal evolution models (Plesa et al., 2018b). Successful models require that the elastic lithosphere thickness values at these two locations are compatible with the present-day estimates. The elastic thickness is anti-correlated to the crustal thickness and surface heat flow (Fig. 1). Regions of thick crust typically associated with the southern hemisphere and in particular with volcanic centers show an elevated heat flow and a thin elastic thickness compared to the northern hemisphere and in particular within impact basins. This is due to the fact that a thicker crust has a higher amount of HPEs and a stronger blanketing effect than a thinner crust. The crustal blanketing effect is produced by the lower crustal conductivity compared to that of the mantle. This leads to higher subsurface temperatures in regions covered by a thick crust compared to areas with a thin crust.

The magnitude of surface heat flow and elastic thickness variations depends on the magnitude of crustal thickness variations. In the following, we discuss the effects of crustal thickness variations for the present-day surface heat flow and elastic thickness pattern taking as examples three geodynamic simulations that use the three different crustal thickness models presented in Fig. 1. The geodynamic models use the bulk heat production rate of Taylor.
and assume that the mantle contains about 43% of the bulk heat production rate, a value that lies in the range suggested by Knapmeyer-Endrun et al. (2021) to produce localized melting regions in the interior at present-day. A crustal thickness dichotomy leads to a dichotomy in surface heat flow and elastic thickness. The smallest surface heat flow and elastic thickness variations are obtained for the thinnest crust scenario (Fig. 1i, g), where crustal thickness variations are more than a factor two smaller compared to the thickest crust scenario. The crustal thickness variations for the case where the crustal density differs across the dichotomy boundary are about 25 km smaller than for the thickest crust scenario. However surface heat flow and elastic thickness variations are more pronounced in this case due to the difference in the crustal density and therefore the amount of crustal HPEs between the southern and northern hemispheres. Due to the lower crustal density of the southern compared to the northern hemisphere, the volumetric heat production in the northern crust is higher than in the southern crust leading to a warmer crust on the northern compared to the southern hemisphere. This is reflected also by the higher surface heat flow in the northern part of the Tharsis region compared to the southern part.

Table 4: Results obtained of the crustal thickness models shown in Fig. 1. All values represent present-day values. $F_s$[min, max] is the average surface heat flux with minimum and maximum values. $T_e$[min, max] is the average elastic thickness with minimum and maximum values calculated assuming a strain rate $\dot{\epsilon} = 10^{-14}$ s$^{-1}$. $F_s^{InSight}$ is the surface heat flux at InSight location. $T_e^{NP}$ is the elastic lithosphere thickness averaged below the north pole ice cap (i.e., within 10° from the north pole), while $T_e^{SP}$ is the elastic lithosphere thickness averaged below the south pole ice cap (i.e., within 5° from the south pole). $T_{CMB}$ is the core-mantle boundary temperature and $F_{CMB}$ is the core-mantle boundary heat flux.

| Output               | thin crust | density dichotomy crust | thick crust |
|----------------------|------------|-------------------------|-------------|
| $F_s$[min, max] [mW m$^{-2}$] | 22.1 (16.3, 30.0) | 22.2 (16.0, 38.9) | 22.3 (14.4, 33.1) |
| $T_e$[min, max] [km] | 267 (185, 326) | 261 (61, 328) | 234 (70, 348) |
| $F_s^{InSight}$ [mW m$^{-2}$] | 20.1 | 22.6 | 19.1 |
| $T_e^{NP}$ [km] | 284 | 288 | 304 |
| $T_e^{SP}$ [km] | 264 | 280 | 236 |
| $T_{CMB}$ [K] | 2094.2 | 2092.7 | 2086.1 |
| $F_{CMB}$ [mW m$^{-2}$] | 2.3 | 2.4 | 2.4 |

The elastic thickness is thickest in areas of thin crust where the interior cools more efficiently. These areas are typically impact basins, with the Hellas impact basin usually recording the highest elastic thickness values. The thinnest elastic thickness is obtained in areas covered by a thick crust, where the presence of an incompetent crustal layer (i.e., a weak crustal layer formed by high crustal temperatures) may decouple the elastic cores of the mantle and the crust, thus further reducing the elastic thickness. This has been suggested to exist at present day in the Tharsis area around Arsia Mons, where the crust is thickest (Grott and Breuer, 2010). Indeed such an incompetent crustal layer is present in the crustal density dichotomy model and the thickest crust scenario. While in the former this is located in Tharsis and in a small area in Elysium, for the latter the incompetent crustal layer is present in the Tharsis area and in smaller locations in the southern hemisphere due to the overall thicker crust in this scenario. As discussed in previous studies, the strongest constraint is given by the elastic thickness at the north pole. While at the south pole, all models present an elastic thickness greater than 150 km being compatible with the latest estimate (Broquet et al., 2021), only the thickest crust scenario presents an elastic thickness larger than 300 km (i.e., 304 km) at the north pole (Table 1), a value that is still lower than the recent estimate of Broquet et al. (2020). A higher elastic thickness may be obtained if the mantle is more depleted in HPEs than assumed in these models. However, a lower heat production in the mantle might lead to scenarios in which partial melt production stops earlier than suggested by the geological record in Tharsis and Elysium (Neukum et al., 2004; Vaucher et al., 2009; Hauber et al., 2011). Another solution to explain the discrepancy between the elastic thickness estimates and the values obtained from geodynamic models would require that the load produced by the polar cap is not yet at elastic equilibrium as discussed by Broquet et al. (2020). This would lead to lower elastic thickness estimates, since in this case the observed deflection would be the sum of a downward deflection caused by the viscous relaxation and an upward deflection caused by some form of postglacial rebound. Whether the north pole is at elastic equilibrium strongly depends on the viscosity of the lithosphere and mantle that is linked to the parameters of the geodynamic model and would require the computation of an individual elastic thickness estimate for each thermal evolution model. Nevertheless, future work needs to address this aspect, since this may significantly affect the number of admissible models that can explain the elastic lithosphere thickness at the north pole of Mars.
In addition to the surface heat flow and elastic lithosphere, the thickness of the crust and its variations can affect the amount and distribution of partial melt that may still be produced in the interior of Mars today. Due to the pronounced crustal blanketing effect and higher amount of crustal HPEs, regions covered by a thick crust can be kept warm and their temperatures can exceed the melting temperature and produce melt up to recent times. Whether melt can still be produced in the martian mantle at present day primarily depends on the amount of mantle HPEs.

The study by Knapmeyer-Endrun et al. (2021) showed that only a limited range of crustal enrichment, i.e., containing between 55–70% of the total bulk of HPEs would lead to localized partial melt production at present day in the interior of Mars. A strong crustal enrichment containing more than 70% of the bulk amount of HPEs would lead to a mantle that is too cold to produce melt at present day. On the other hand a mantle containing more than 45% of the bulk amount of bulk amount of HPEs would be too warm and lead to wide-spread melting at present day. For a crust with an average thickness at the upper end of values obtained from InSight’s seismic data this indicates a crustal enrichment in thorium and potassium similar to the surface abundance as measured by the gamma-ray instrument on board Mars Odyssey (Hahn et al., 2011; Taylor et al., 2006a). A thinner crust, on the other hand, requires an enriched component in the subsurface in order to avoid wide-spread melting in the interior of Mars at present day (Knapmeyer-Endrun et al., 2021).

In Fig. 2 we show the present-day melt fraction and the depth of the melt zone for the three models presented in Fig. 1. The melting temperature is taken from Ruedas and Breuer (2017), who updated the solidus parametrization of Ruedas et al. (2013) to include more recent melting experiments from Collinet et al. (2015) and Matsukage et al. (2013). Furthermore, Ruedas and Breuer (2017) include a correction to account for the effects of Na, K, and Ca as suggested by Kiefer et al. (2015) that leads to about 35 K lower solidus for the primitive martian mantle compared to the terrestrial mantle. Additionally, the solidus used in each model in Fig. 2 considers the effect of mantle depletion due to crust formation. The solidus is increased linearly with the degree of depletion that each model experienced according to the crustal volume.

The least melt is produced at depths larger than 350 km in the thin crust model (Fig. 2a and d) that also has the smallest crustal thickness variations. The crustal blanketing effect is more pronounced for the other two models using a crustal density dichotomy and a higher density crust, respectively, due to a locally thicker crust in these models. This results in a larger amount of partial melt (higher melt fractions and more melt regions) and shallower melt zones compared to the thin crust model.

In general, geodynamic models have difficulties to produce partial melt zones at present-day beneath the Elysium volcanic province, even though mantle plumes are present there. Melting takes place mostly in Tharsis and underneath the southern hemisphere, as these areas are typically covered by a thicker crust. Since the Elysium province lies in the northern lowlands, it is difficult to focus mantle plumes underneath it that are hot enough to
produce melt at present day. Interestingly, the crustal density dichotomy model shows melting zones focused in Tharsis and Elysium. The melting zone in Elysium is likely due to a thicker crust in this region compared to the southern hemisphere and due to the reduced differences in crustal thickness between the northern and southern hemispheres caused by using a lower crustal density in the south compared to the north. We note, however, that the density difference between the southern and the northern crust is quite extreme in this model (i.e., 400 kg m\(^{-3}\)), and whether a crustal thickness model with smaller density variations between north and south can produce a partial melting zone at present day beneath Elysium needs to be tested by future studies.

4 Mantle dynamics, seismic velocities variations, and tidal dissipation

The thermal state of the mantle, the thermal lithosphere thickness, as well as the location of hot mantle plumes and cold downwellings can affect the variation of seismic velocities. Typically, the thickness of the thermal lithosphere is the sum of the stagnant lid thickness and of the thermal boundary layer where convective instabilities initiate. Below the thermal lithosphere, the mantle temperature usually follows an adiabatic profile. However, deviations from an adiabatic temperature profile may occur if convection is sluggish due to a high pressure-dependence of the viscosity or due to strong cooling of the interior, in which case the average temperature profile lies between an adiabatic and a conductive profile (Fig. 3). Parametrized thermal evolution models typically use either an adiabatic mantle temperature profile or a conductive one, if convection stops (i.e., the Rayleigh number drops below a critical value). The 2D/3D geodynamic models, on the other hand, self-consistently calculate the thermal profile, and in these models, depending on the rheological parameters, the thermal profile in the mantle may lie between an adiabatic and a conductive profile.

Figure 3: Effect of the pressure dependence of the viscosity, which is given by the activation volume, on the temperature (a) and viscosity (b) profiles at present day. The values for the activation volume are taken from [Hirth and Kohlstedt (2003)].

The thermal state of the interior is to first order affected by the mantle viscosity. However, rheological parameters such as the activation energy and activation volume that are determined by laboratory deformation experiments have large uncertainties. While the activation energy of olivine aggregates was measured to lie at about 375 ± 75 kJ mol\(^{-1}\) for diffusion and 520 ± 40 kJ mol\(^{-1}\) for dislocation creep (Hirth and Kohlstedt, 2003), the activation volume is one of the most poorly constrained parameters with values between 0 and 20 cm\(^3\) mol\(^{-1}\) (Hirth and Kohlstedt, 2003). These values will substantially affect the mantle temperature profile (Fig. 3) and also the mantle convection pattern. An activation volume of 6 cm\(^3\) mol\(^{-1}\) leads to a larger number of plumes and downwellings compared to an activation volume of 10 cm\(^3\) mol\(^{-1}\), as illustrated in the temperature maps at mid-mantle depth (Fig. 4a and b). However, further increasing the activation volume would lead to scenarios, in which convection in the lower part of the mantle is weak or even absent. This reduces the thickness of the convective layer and increases the wavelength of the convection pattern (Fig. 4c and d).

The viscosity could also be affected by the presence of water in the mantle. Water concentrations in excess of 100...
Figure 4: Temperature variations at mid-mantle depth (775 km depth) corresponding to the temperature profiles shown in Fig. 3. The models use an activation volume of 6 cm$^3$ mol$^{-1}$ (panel a), 10 cm$^3$ mol$^{-1}$ (panel b), 15 cm$^3$ mol$^{-1}$ (panel c), and 20 cm$^3$ mol$^{-1}$ (panel d). The color scale has been clipped to show the locations of mantle plumes (bright colors) and downwellings (dark colors). For orientation the 0 km level of the surface topography is indicated by white contour lines.
ppm can reduce the viscosity by about two orders of magnitude (Karato and Wu, 1993) and can significantly affect mantle cooling. On Mars, recent petrological analyses of martian meteorites suggest a bulk water content of 137 ppm, with crustal abundances of 1410 ppm and mantle water contents between 14 and 72 ppm (McCubbin et al., 2016). A wet mantle rheology during most of the thermal history would not be able to reproduce strong mantle plumes in recent times (Plesa et al., 2018b) as required by the petrological evidence for local mantle temperatures (Filiberto and Dasgupta, 2015; Kiefer and Li, 2016). In addition, a dry rheology was favored by models that coupled the thermal and orbital evolution of Mars and its moon Phobos, in order to reproduce the orbital evolution of Mars’ closest satellite (Samuel et al., 2019). Thus, according to thermal evolution models, water in the martian mantle was most likely lost during the earliest planetary evolution and most of the thermal history was characterized by a dry mantle rheology. We note, however, that geochemical reservoirs may complicate this interpretation, as they could trap water in isolated regions inside the mantle and lithosphere (Breuer et al., 2016).

The present-day thermal state of the interior is the result of billion years of thermal evolution. Mantle plumes and cold downwellings may be present in the interior of Mars at present day and would lead to temperature variations in the interior and to variations in the thermal lithosphere thickness. These effects can only be investigated by using 2D and 3D geodynamic models. Previous models showed that the thermal lithosphere can be substantially thinner at the location of mantle plumes (Kiefer and Li, 2009), as their higher temperature decreases locally the viscosity and allows the silicate material to flow (i.e., to convect) at shallower depths. Conversely, the thermal lithosphere is thicker above cold mantle downwellings. Additionally, the variations of the thermal lithosphere thickness are affected by the crustal thickness variations, as the latter has a higher amount of HPEs and a lower conductivity compared to the mantle. This leads to a higher lithospheric temperature and thinner lithospheric thickness beneath areas covered by a thick crust (e.g., beneath volcanic provinces) compared to regions of thin crust (e.g., impact basins). For models that include crustal thickness variations, the largest variations in temperature are observed in the lithosphere. We also note that the largest temperature variations are obtained for models with a thick crust that exhibits larger crustal thickness variations. In Fig. 5a we show average temperature profiles and corresponding temperature variations at present day for the three thermal evolution models presented in Fig. 3. The model with a crustal density dichotomy between the northern and southern hemisphere and the model with a crustal density of 3000 kg m⁻³ have crustal thickness variations of 127.2 km and 152.3 km, respectively, that lead to larger lithospheric temperature variations than the crustal thickness model with a crustal density of 2550 kg m⁻³, which has a difference of only 60.8 km between the minimum and maximum crustal thickness values.

![Figure 5](image)

**Figure 5:** Profiles of the temperatures (panel a) as well as the shear wave velocities (panel b) and compressional wave velocities (panel c) for the models presented in Fig. 3. The TAY13 (Taylor, 2013) was assumed for the seismic velocities calculation. Full lines indicate the average profiles throughout the mantle, while the shaded regions show the corresponding variations.

On Mars, the average thermal lithosphere thickness has been estimated based on the evaluation of seismic events recorded by InSight. Using direct and surface-reflected body wave phases, a thermal lithosphere thickness of 400–600 km is required to explain the differential travel times obtained for seismic events at epicentral distance between 25° and 75° from InSight location and with moment magnitudes between 3 and 4 (Khan et al., 2021). InSight’s estimate of the average thermal lithosphere thickness of Mars is thicker than the thermal lithosphere on the Earth and suggests that Mars has significantly cooled during its thermal history. This thick lithosphere and its thermal gradient inferred from InSight data control the formation of low-velocity zones in the interior of Mars that have been proposed in previous studies (Mocquet and Menvielle, 2000; Zheng et al., 2015). Additionally, a recent study
Figure 6: Effects of mantle composition on the compressional wave velocities (panels a, b, and c) and shear wave velocities (panels d, e, and f) for three different mantle compositions: TAY13 [Taylor, 2013], YMD20 [Yoshizaki and McDonough, 2020], and KSLRD22 [Khan et al., 2022]. Panels a and d show the thin crust end-member case, panels b and e present the results for the density dichotomy crust, and panels c and f show the thick crust end-member case.
by Plesa et al. (2021) showed that 3D thermal evolution models with a crust containing less than 20% of the bulk amount of HPEs would lead to a hot interior and a thin lithosphere. These models are incompatible with InSight observations, as they would lead to S-wave shadow zones for high-quality events in the Cerberus Fossae region, for which clear P- and S-waves arrivals were recorded.

Temperature variations affect the seismic velocities with the strongest velocities variation being present in the lithosphere and at the depth of the olivine to wadsleyite phase-transition (Fig. 5 and c). We note that the effects of composition are minor compared to the effect of temperature variations in the lithosphere. In Fig. 6 we show the differences between the average seismic velocities profiles for the three models presented in Fig. 5. We tested three of the most recent compositions that have been proposed for Mars: the TAY13 composition (Taylor 2013), the YMD20 composition (Yoshizaki and McDonough 2020), and the KSLRD22 composition (Khan et al. 2022). While the seismic velocities have been computed using these compositions, the bulk amount of HPEs that was used in all thermal evolution models was taken from TAY13. Other HPEs models with a higher abundance of radiogenic elements such as the model by Yoshizaki and McDonough (2020) would require a higher crustal enrichment in HPEs and a similar amount of mantle HPEs as the models shown here in order to avoid wide-spread melting in the mantle at present day. Thus, even for other HPEs models the mantle temperature would be similar to the profiles shown in Fig. 5.

For all three compositional models tested here, the seismic velocities are nearly identical in the upper mantle and show slight differences in the lower mantle, with minimally lower seismic velocities values for the YMD20 and KSLRD22 compositions mainly due to the lower FeO content of these models (14.7 ± 1.0 wt% for YMD20 and 13.7 ± 0.4 wt% for KSLRD22 compared to 18.1 ± 2.2 wt% for TAY13).

The temperature and hence the seismic velocities in the lithosphere follow the crustal thickness variations. The crustal thickness pattern controls their variations down to a depth of 400 km or even deeper in particular for models with a thick crust (Fig. 7). In Fig. 7 the thin crust model (average crustal thickness of 40.6 km, left column) shows a pattern of the S-wave velocity variations that closely follows the crustal thickness pattern at 150 km depth. Lower-than-average seismic velocities are observed below the southern hemisphere and are caused by the warmer temperatures due to a thicker crust compared to the northern hemisphere. Conversely, the areas covered by a thin crust, i.e., the northern hemisphere and large impact basins such as Hellas, show seismic velocities higher than the average value, due to the more efficient cooling and hence colder temperatures than those beneath the southern hemisphere. This seismic velocities pattern is no longer visible at 400 km depth. For the thick crust model (average crustal thickness of 71.4 km, right column), however, a dichotomy in the S-wave velocity variations is still visible at 400 km depth. Since all models in Fig. 7 use the same bulk heat production and the same amount of HPEs in the mantle, the difference is caused by the stronger crustal thickness variations and the more pronounced blanketing effect due to the low crustal conductivity in the thick crust scenario compared to the thin crust case. In the crustal density dichotomy case (middle column), the crustal thickness pattern is more complex than in the previous two models, but the Tharsis region is clearly distinguishable on the map of S-wave velocity variations at 400 km depth.

Seismic velocities variations are small and about 1% in the convecting mantle. Larger variations are observed again closer to the CMB, where negative seismic velocity gradients may be locally present due to the stability of larger proportions of garnet and ferropericlase at the expense of ringwoodite (Plesa et al. 2021). We note however that this depends on the chosen mineralogical model and mostly appears for the Taylor-composition (Taylor 2013), but is absent for the Yoshizaki- and Khan-compositions (Yoshizaki and McDonough 2020, Khan et al. 2022).

Depending on the location of seismic events, the propagation of their seismic waves will encounter not only a different crustal thickness on the path to the seismic station, but also a different lithospheric thickness and lithospheric temperature. In Fig. 8 we show the differences between seismic velocities profiles at present-day at three different locations on Mars (Tharsis, Utopia, and InSight) for the three thermal evolution models presented in Fig. 4 and compare them with their corresponding average profiles. The differences in the uppermost 400 km can be substantial. The uppermost differences are observed between Tharsis volcanic province (profile taken at −115° longitude and 0° latitude) and Utopia impact basin (115° longitude and 45° latitude), and are most extreme for models with a thick crust in the Tharsis area (the thick crust model and the model with a different density between the northern and southern hemispheres). Interestingly, the model with a different density between the northern and southern hemispheres shows the largest variations. This is caused by the higher volumetric heat production in the northern hemisphere compared to the southern hemisphere in this model. On the other hand, the average profile and the profile at the InSight landing site are nearly identical. The largest difference between InSight and average profiles is observed for the thickest crust scenario, where crustal thickness variations between these two locations are more pronounced than in the other two crustal thickness models.

The uppermost layers are dominated by the seismic velocities of the crust. While for the individual profiles at the three locations a sharp transition occurs at the crust mantle interface, this transition is more gradual for
Figure 7: Seismic velocities variations at different depths throughout the mantle calculated for TAY13 composition. Left column (panels a, d, g and j) shows the thin crust end-member case. Middle column (panels b, e, h, and k) presents the density dichotomy crust model, and the right column (panels c, f, i, and l) shows the thick crust end-member case. For orientation, the gray contour lines show the 0 km level of surface topography.
Figure 8: Seismic velocities profiles at selected locations compared to the average profiles. InSight profile is drawn at 136° longitude and 4° latitude. Tharsis profile is taken at −115° longitude and 0° latitude, while the Utopia profile was selected at 115° longitude and 45° latitude. Panels a and d show the seismic velocities profiles for the thin crust end-member, panel b and e for the density dichotomy crust case, and panels c and f for the thick end-member case.
the average profile. This is due to the fact that for the average profiles both mantle and crustal areas are present at the same depth. In the uppermost layers the crustal areas dominate but with deeper depth they become smaller being replaced by mantle areas. Right below the crust, the seismic velocities reflect the large variations of the lithospheric temperatures, while deeper in the convecting mantle these variations become much smaller and therefore the difference in seismic velocities is minor. The olivine to wadsleyite phase transition is clearly visible for all models. The depth of the phase transition depends on the temperature, and for the Tharsis profile, which has a higher local temperature, the average phase transition depth is at about 1110 ± 10 km, depending on the thermal evolution model. For the InSight profile the average phase transition depth is shallower and about 1070–1080 km, due to a lower temperature at this location. All models show the same average depth for the olivine to wadsleyite phase transition at ∼1095 km for the average profile, since the latter is nearly identical for all models (cf. Fig. 5).

In addition to the seismic velocities, the thermal state of the mantle together with the size and state of the core (solid or liquid, see Section 5) also affects the tidal deformation of a planet. The latter has been determined for Mars from radio tracking measurements from Mars Odyssey, Mars Reconnaissance Orbiter, and Mars Global Surveyor (Konopliv et al., 2016; Genova et al., 2016; Konopliv et al., 2020). The lag of the tidal deformation caused by Mars’ closest moon, Phobos, is given by the phase lag ε that describes the dissipation inside Mars and is linked to the thermal state of the interior through the viscosity (Nimmo and Faul, 2013). The dissipation can be also expressed in form of the tidal quality factor Q that is defined as $1/\sin(\varepsilon)$. Low values of Q would indicate a dissipative mantle caused by a low viscosity and hence high mantle temperature, whereas a cold interior and consequently a high viscosity would lead to high Q values.

On Mars, current available estimates for the tidal quality factor Q calculated at the main tidal period of Phobos (5.55 hours) range between 72 and 105 (Ray et al., 2001; Lainey, 2016) and indicate a more dissipative interior than that of the Earth, for which a tidal quality factor of 280 was calculated at the lunar semidiurnal terrestrial tide (Ray et al., 2001). Here we recalculate the tidal quality factor using the approach from Zharkov and Gudkova (2005) that was also used by Khan et al. (2018):

$$Q \approx \frac{559}{k_2}$$

Using the latest $k_2$ estimate of 0.174 ± 0.008 (Konopliv et al., 2020), we find a tidal quality factor of 97.3 ± 4.5. However, the error bars would increase when accounting for the frequency-dependency of $k_2$, higher tidal terms, and the fact that dissipation occurs in both Mars and Phobos. Thus, following the approach of Khan et al. (2018) we increase the error bars and use a tidal quality factor $Q$ of 97 ± 12, a range that includes the previous estimates of Khan et al. (2018) as well as the new values obtained from the most recent $k_2$ estimate.

Previous studies have used prescribed thermal profiles (Nimmo and Faul, 2013) or temperature profiles from mantle convection models (Plesa et al., 2018b) to calculate the dissipation in the interior of Mars and compare the results with observations. The study by Nimmo and Faul (2013) uses an extended Burgers model for dry olivine and finds that, for a grain size of 1 cm, a present-day potential temperature of 1625 ± 75 K is required to explain the dissipation in the interior of Mars. Plesa et al. (2018b) used the present-day thermal state from mantle convection models and computed the tidal quality factor $Q$. Using $Q$ estimates in the range of 99.5 ± 4.9 (Konopliv et al., 2016; Lainey, 2016), this study concluded that models with an inefficient cooling of the interior caused by either a high amount of HPEs in the mantle or a large increase of the viscosity with depth (i.e., due to a high activation volume) would be too dissipative to satisfy the constraints. Conversely, models that contain nearly all HPEs in the crust and have a cold present-day mantle lead to a much lower dissipation than the suggested values for Mars. Thermal evolution compatible with the $Q$ estimates of Lainey (2016) indicates that between 37.6% and 68.3% of the bulk amount of HPEs are concentrated in the crust (Plesa et al., 2018b).

Here, we calculate the tidal deformation of the three thermal evolution models presented in Fig. 1. These models contain 57% of the total bulk amount of HPEs in their crust, and thus lie within the range of models that were found compatible with $Q$ estimates by Plesa et al. (2018b). To compute the tidal deformation, we use a semianalytical model based on the normal mode theory for radially stratified viscoelastic bodies (Sabadini and Vermeersen, 2004). The results are shown in Fig. 9 and Table 5 and are discussed in detail below.

The model uses 100 layers for the mantle and 1 layer for the core, which is assumes to be homogeneous. It uses as inputs the density, viscosity and rigidity. While the viscosity profile comes from the geodynamic simulations, the density and rigidity profiles depend on the mineralogical model and are calculated by the thermodynamic code Perple_X (Connolly, 2009) using the thermodynamic formulation and database of Stixrude and Lithgow-Bertelloni (2011). All three profiles (i.e., viscosity, density, and rigidity) are temperature and pressure-dependent and are calculated from the temperature profiles obtained by the geodynamic models.

For the tidal deformation calculations, we used the Andrade rheological model and assumed that the planet is incompressible. Nevertheless, we note that the assumption of compressibility might render corrections to $k_2$ that
Figure 9: Panel a: Viscosity profiles throughout the mantle for the models shown in Fig. 1. Panel b and c: mantle density and rigidity profiles, respectively, for the thick crust end-member using three different mantle compositions TAY13, YMD20 and KSLRD22. Panel d: calculated tidal quality factor $Q$ and tidal Love number $k_2$ for the three mantle models (panel a) and three compositions (panel b and c).
Table 5: Tidal deformation results for the three models presented in Fig. 1 using an incompressible Andrade model, three different mantle mineralogies, and various values of the tidal parameter $\alpha$ used in the Andrade model. According to Castillo-Roge et al. (2011) the tidal parameter $\zeta$ was set to 1 for all calculations.

| Tidal parameter | thin crust density dichotomy crust thick crust | TAY13 composition (Taylor 2013) | YOS20 composition (Yoshizaki and McDonough 2020) | KHA22 composition (Khan et al. 2022) |
|-----------------|---------------------------------------------|---------------------------|---------------------------------------------|---------------------------------------------|
| $\alpha = 0.1$  | $Q$ 0.191 $k_2$ 56.50 0.191 | $Q$ 0.191 $k_2$ 56.60 0.191 | $Q$ 0.191 $k_2$ 56.81 0.192 |
| $\alpha = 0.15$ | 81.40 0.178 | 81.71 0.178 | 82.47 0.179 |
| $\alpha = 0.2$  | 137.94 0.173 | 138.75 0.173 | 140.92 0.173 |
| $\alpha = 0.25$ | 253.90 0.171 | 255.91 0.171 | 261.50 0.171 |
| $\alpha = 0.3$  | 489.38 0.170 | 494.19 0.170 | 508.01 0.170 |
| $\alpha = 0.4$  | 1955.25 0.169 | 1981.69 0.169 | 2060.93 0.170 |
| $\alpha = 0.1$  | $Q$ 0.191 $k_2$ 56.76 0.191 | $Q$ 0.191 $k_2$ 56.86 0.191 | $Q$ 0.191 $k_2$ 57.09 0.192 |
| $\alpha = 0.15$ | 81.90 0.179 | 82.21 0.179 | 83.00 0.179 |
| $\alpha = 0.2$  | 138.98 0.173 | 139.81 0.173 | 142.03 0.174 |
| $\alpha = 0.25$ | 256.20 0.171 | 258.24 0.171 | 263.93 0.172 |
| $\alpha = 0.3$  | 494.50 0.170 | 499.37 0.170 | 513.42 0.171 |
| $\alpha = 0.4$  | 1980.81 0.169 | 2007.58 0.170 | 2088.06 0.170 |
| $\alpha = 0.1$  | $Q$ 0.191 $k_2$ 56.69 0.189 | $Q$ 0.191 $k_2$ 56.79 0.189 | $Q$ 0.191 $k_2$ 57.01 0.190 |
| $\alpha = 0.15$ | 81.77 0.177 | 82.08 0.177 | 82.87 0.177 |
| $\alpha = 0.2$  | 138.72 0.171 | 139.53 0.171 | 141.74 0.172 |
| $\alpha = 0.25$ | 255.58 0.169 | 257.60 0.169 | 263.27 0.170 |
| $\alpha = 0.3$  | 493.03 0.168 | 497.87 0.168 | 511.85 0.169 |
| $\alpha = 0.4$  | 1972.59 0.168 | 1999.18 0.168 | 2079.16 0.168 |
are of the order of the observational uncertainty. While other rheological models exist and have been applied to calculate the tidal deformation of Mars and other rocky planets, the advantage of the Andrade model is the small number of parameters it requires for the calculations (Castillo-Rogez et al., 2011; Efroimsky, 2012). A detailed review of the theory of viscoelasticity and tidal response, as well as their application to constrain the interior structure of Mercury, Venus, Mars, the Moon, and icy satellites is given in (Bagheri et al., 2022).

A recent study by Bagheri et al. (2019) has compared various rheological models for Mars and found that all models can fit the observations when using a single frequency (i.e., at the main period of Phobos), but information of the dissipation at additional frequencies could help to distinguish between the current rheological models. The same study concluded that the Maxwell rheology would require very low viscosities to fit the available data and rheologies such as Andrade, extended-Burgers, or Sundberg-Cooper are more appropriate to use when studying tidal dissipation.

The Andrade model that is used here is able to describe all components of deformation (elastic deformation, viscous creep, and the transient Andrade creep) and requires in total four parameters: the viscosity, the rigidity and two empirically determined parameters $\alpha$ and $\zeta$. The parameter $\zeta$ describes the ratio between the timescales of the anelastic Andrade creep and the Maxwell body and has been found to be close to one (Castillo-Rogez et al., 2011). On the other hand $\alpha$ describes the duration of the transient response, and values for olivine-rich mantle rocks lie between 0.1 and 0.5, and mostly between 0.2 and 0.4 (Castillo-Rogez et al., 2011).

The average thermal state of the geodynamic models, which was used to calculate the tidal deformation in Fig. 9, is very similar (cf. Fig. 5a). Although the temperature and seismic velocities variations in the lithosphere can be significantly different between the three geodynamic models (cf. Fig. 7), the average viscosity, density, and rigidity profiles are very similar (Fig. 9a, b, and c), with TAY13 composition showing slightly larger densities and higher rigidities than YOS20 and KHA21 composition, due to the higher FeO content. Thus, the tidal dissipation values mainly depend on the chosen value for $\alpha$, for which a value between 0.2 and 0.15 seems to fit best the observed dissipation in the interior of Mars (Fig. 9).

5 Core radius estimates and their implications for the interior dynamics

The core of a terrestrial planet is a witness of the earliest planetary differentiation, when metal and silicates separate to form the layers inside the planet. The size of the core is essential to determine the thickness of the silicate layer (mantle and crust), when knowing the planet’s radius. The thickness of the silicate layer in turn affects the mantle flow and the convection pattern (i.e., number of mantle plumes and their distribution). The latter can be linked to surface geological features such as volcanic and tectonic provinces. For a detailed review that describes the methods and the progress in determining of the core size of the Earth, Mars, and Moon, as well as future opportunities for terrestrial exoplanets we refer the reader to (Knapmeyer and Walterová, 2022).

Many geodynamic studies have investigated the formation of the martian crustal thickness dichotomy, proposing a degree-one or ridge like convection pattern. This pattern is largely favored for models using a small core that allows for the presence of an endothermic phase transition at the base of the mantle (Harder and Christensen, 1996; Breuer et al., 1998), similar to the 660-phase transition on the Earth. Models employing a specific mantle viscosity structure with a viscosity increase in the mid-mantle (Zhong and Zuber, 2001; Roberts and Zhong, 2006; Keller and Tackley, 2009) and models that included the combined effects of a giant impact and the subsequent dynamics in the mantle (Golabek et al., 2011) were also able to produce a degree-one mantle pattern, but in these cases too the core radius was about half of the planetary radius or smaller.

InSight’s measurements have revealed that the martian core has a radius of 1830 ± 40 km and is more than half the planet’s radius (Stähler et al., 2021). This is consistent with estimates of the tidal Love number $k_2$ (Konopliv et al., 2016; Genova et al., 2016) that were previously combined with thermal evolution models and suggested a core radius strictly larger than 1800 km (Plesa et al., 2018b). The $k_2$ value of Konopliv et al. (2016) of 0.169 ± 0.006 has been recently updated by Konopliv et al. (2020) to 0.174 ± 0.009. The most recent estimate was corrected for atmospheric tides, and the uncertainties account for the fact that the correction for atmospheric tides depends on the atmospheric conditions at the time of observations (Konopliv et al., 2020). As shown in Fig. 9, models using a core radius of 1850 km, a value that was previously used by Plesa et al. (2018b) and is consistent with the seismic detection of the martian core (Stähler et al., 2021), are able to fit the latest $k_2$ estimate of Konopliv et al. (2020).

The large size of the core excludes the possibility of having a bridgmanite-dominated lower mantle on Mars (Stähler et al., 2021), as it is the case for the Earth. An endothermic phase transition at the base of the martian
mantle will no longer occur, as the pressure is too low for this to take place. Even for models with smaller core radii (1700 – 1360 km) an endothermic phase transition at the base of the present-day martian mantle was only marginally possible requiring CMB temperatures in excess of 2100 K (Spohn et al. 1998). In addition, the large radius of the core leads to a small scale convection pattern with many small plumes distributed throughout the mantle as illustrated in Fig. 10.

All three models in Fig. 10 have been built similarly to the models presented in (Plesa et al. 2016). They use the same crustal thickness with a crustal density of 2800 kg m$^{-3}$ and a crustal enrichment that matches the average value derived from GRS measurements. The crustal thickness is derived from gravity and topography data and matches the crust-mantle discontinuity at InSight landing site, as observed in the seismic measurements (Knapmeyer-Endrun et al. 2021; Wieczorek et al. 2022). The radius of the core has been varied between the three models with values of 1500 km, 1700 km, and 1850 km. While the former two values are incompatible with the recent InSight data, the last value of 1850 km lies well within the current core radius estimates (Stähler et al. 2021).

The models show a shallow subsurface that is mostly dominated by the crustal thickness variations with a temperature pattern similar to the crustal thickness pattern (Fig. 10h–f). The effects of mantle plumes are more pronounced for 1500 km core radius compared to the 1850 km core radius. It can be observed that the case with the largest core shows a short wave-length convection pattern compared to the small core model. This is illustrated both in the temperature maps at mid-mantle depth (Fig. 10g–i) and at 100 km above the CMB (Fig. 10j–l) that show the presence of a larger number of plumes and downwellings for the model with a core radius of 1850 km compared to the case with a core radius of only 1500 km.

A large core that leads to a small scale mantle convection pattern is at odds with the formation of the martian crustal thickness dichotomy through an endogenous process. However, exogenic processes such as the sequence of one or several large impacts may represent a key mechanism (Wilhelms and Squyres 1984; Frey and Schultz 1988). Such scenario is compelling, as it has been proposed to explain the elliptical nature of the dichotomy (Andrews-Hanna et al. 2008; Nimmo et al. 2008; Marinova et al. 2008). The effect on the subsequent dynamics in the mantle may lead to mantle plumes in Tharsis and Elysium that with time may be stabilized by the insulating effect of a thicker crust at those locations (Schumacher and Breuer 2006). This scenario, however, would not exclude weaker thermal anomalies at other locations, that may have led to shorter episodes of volcanic activity (e.g., Syrtis Major, the Circum Hellas province) or may have never resulted in the buildup of volcanic provinces at the surface.

Another important aspect of a large martian core is that it requires a high amount of light elements to be able to match the planet’s mass. Sulfur cannot be the only light element in the core, as its amount would then exceed the abundance found in EH-chondrites, the most sulfur rich building blocks. Thus, other elements such as oxygen, carbon, and hydrogen are required to match both the core density and to be compatible with geochemical arguments (Stähler et al. 2021; Khan et al. 2022 and references therein). The composition of the martian core has major consequences for its evolution and the generation of an early magnetic field. A large amount of light elements in the core could lead to a scenario in which the core crystallizes from the top down (iron snow) by forming iron particles at the top of the core due to a steeper melting temperature than the core adiabatic profile (Stewart et al. 2007; Rivoldini et al. 2011; Breuer et al. 2015; Helffrich 2017; Davies and Pommier 2018; Hemingway and Driscoll 2021).

A recent study by Hemingway and Driscoll (2021) investigated the crystallization of the martian core and found that Mars may possess a partially solid core today. However, the large amount of light elements places the core composition close to the eutectic and likely prevented the crystallization of an inner core (Stähler et al. 2021), due to the significant decrease of the core melting temperature (Mori et al. 2017). In addition, Hemingway and Driscoll (2021) used a simple thermal evolution model without explicit treatment of the stagnant lid evolution. This most likely underestimates the mantle and core temperatures. Typical stagnant lid thermal evolution models (e.g., Breuer and Spohn 2003; Morschhauser et al. 2011; Plesa et al. 2018b; Samuel et al. 2019, 2021) suggest that even at present day the CMB temperature is too high to allow for core crystallization, in particular for a high mantle viscosity that was found compatible with additional geophysical and seismic constraints (Plesa et al. 2018b; Samuel et al. 2019; Knapmeyer-Endrun et al. 2021). However, core crystallization would undoubtedly take place in the future, when the core temperature has sufficiently decreased to allow for core crystallization. The exact time and style of crystallization will strongly depend on the core composition and thermal state (Stewart et al. 2007).

The early martian dynamo was most likely driven by thermal buoyancy inside the core until at least 3.7 Gyr ago (Mittelholzer et al. 2020), placing important constraints on the heat flow at the CMB. Thermal evolution scenarios that maintain a CMB heat flow above the critical core heat flow, above which thermal convection in the core sets in, for at least 800 Myr need to be investigated in future studies. An important step for answering this question has been undertaken by Greenwood et al. (2021), who used 1D parametrized thermal evolution models and showed
Figure 10: Mantle convection pattern (panels a, b, and c) and temperature variations throughout the mantle. All models use a crustal thickness with an average value of 61.3 km and a crustal density of 2800 kg m$^{-3}$. Left column (panels a, d, g, and j) shows a model with a core radius of 1500 km. Middle column (panels b, e, h, and k) shows a model with a 1700 km core radius, while right column (panels c, f, i, and l) presents a model with a core radius of 1850 km. Only the model shown in the right column is compatible with core radius estimates by InSight.
that a prolonged thermally driven dynamo can be sustained. Successful models require core thermal conductivities in the range of $16 - 35 \, \text{W m}^{-1} \, \text{K}^{-1}$ and mantle reference viscosities of $10^{21}$ Pa s or smaller and activation volumes smaller than 6 cm$^3$ mol$^{-1}$ (Greenwood et al., 2021). However, geodynamic models in a 3D geometry require higher viscosities and/or activation volumes to explain localized melting at recent times (Knapmeyer-Endrun et al., 2021) and a large present-day elastic lithosphere thickness at the north pole of Mars (Plesa et al., 2018b). Thus future models need to investigate if both constraints for the early magnetic field and recent thermal state can be matched.

6 Seismogenic layer thickness and the present-day seismicity

Seismic observations provide the most direct view into the interior of a planetary body and reveal the level of activity that the planet experiences. While seismic observations have greatly improved our understanding of the interiors of the Earth, Moon, and Mars, currently, the seismic activity for other planetary bodies such as Mercury, Venus, or icy satellites can only be indirectly estimated with large uncertainties. A comprehensive review of the current state of knowledge of planetary seismology and directions for future seismic investigations of planetary bodies is given in Stähler and Knapmeyer (2022).

In the absence of plate tectonics and because Mars is smaller than the Earth, its seismicity was suggested to be lower than that of the Earth and mainly driven by planetary cooling. However, the level of seismic activity was expected to be larger than the seismicity of the Moon recorded by the Apollo seismic measurements (Ewing et al., 1971; Toksöz et al., 1974). Previous seismic measurements on Mars were performed by Viking in the 1970s (Anderson et al., 1976). While the seismometer on Viking 1 failed to uncage and could not record any data, Viking 2 collected data between 1976 and 1978. However, these measurements are strongly contaminated by noise caused by lander vibrations due to wind, given the location of the seismometer on the lander deck. During 146 sols of operation only one event recorded by Viking 2 seismometer could be interpreted as a marsquake (Anderson et al., 1977), but a seismic origin is difficult to establish in the absence of wind data during the event.

In the absence of unambiguous seismic recordings from Mars, previous studies have estimated the level of seismic activity based on the analysis of surface faults (Golombek et al., 1992; Golombek, 1994, 2002) and from numerical models of planetary cooling (Phillips, 1991; Knapmeyer et al., 2006; Plesa et al., 2018a). Maps of tectonic centers were compiled based on orbital imaging of the surface. Using the Mars Orbiting Laser Altimeter shaded topographic relief maps, Knapmeyer et al. (2006) compiled a global fault catalog and used it to predict the martian seismicity and the distribution of epicenters by associating the event size with fault length. Another study by Plesa et al. (2018a) used 3D geodynamic thermal evolution models combined with spatial variations of crustal thickness to evaluate the seismogenic layer thickness and the present-day martian seismicity. While all previous models predicted that Mars is seismically active today with a seismicity between that of the Moon and that of the Earth, the uncertainties of the annual seismic moment covered several orders of magnitude.

Since more than three years, InSight’s seismometer (SEIS) has been recording seismic events on Mars. In the absence of microseismic events that are observed on the Earth, SEIS is able to record extremely small amplitude events on Mars (Lognonné et al., 2020). Although sensitive to the martian wind that leads to a noisy environment during the martian mid-day (Giardini et al., 2020), SEIS was able to record over 2000 teleseismic events (InSight Marsquake Service, 2022) mostly during the late afternoon and evening, when the noise level is low. For some of these events the location could be determined and several of them have been localized in the Cerbeus Fossae region (Zenhäusern et al., 2022) – a young fault system with a minimum age of 10 Myr situated between 20° and 40° east of the InSight landing site (Taylor et al., 2013). A recent study by Horleston et al. (2022) reported on distant seismic events, one of which could be located in Valles Marineris (146°±7°). For many events, however, a localization is difficult in particular due to large uncertainties in the backazimuth that are caused by high scattering and noise levels in the seismic data. Therefore, for many of the recorded seismic events only a distance can be provided, while the direction that is required to determine the location of the source remains unclear (Giardini et al., 2020). Recent advances in the study of these events, by using a comprehensive polarization analysis, have been applied to improve the estimates of the distribution of seismicity on Mars (Zenhäusern et al., 2022).

On Mars planetary cooling was thought to be the main source of present day seismicity. However, high frequency events detected by InSight may be driven by solar illumination, the CO$_2$ cycle or annual solar tides (Knapmeyer et al., 2021). Moreover, the high level of seismicity observed in Cerberus Fossae (Zenhäusern et al., 2022) could be indicative of processes such as magma ascent through the crust and lithosphere that may be ongoing on Mars. Indeed, some of the low frequency marsquakes have been suggested to be related to volcanic tremor in Elysium Planitia region (Kedar et al., 2021). Thus, seismicity may not only be linked to the cooling of the interior, but also to ongoing magmatic processes, and knowledge about the level of seismic activity and location of seismic events can help to constrain the evolution and present-day state of the martian mantle. Global thermal evolution models which
use crustal thickness variations derived from gravity and topography data and anchored by the seismic observations at InSight landing site, show a close correlation between the crustal thickness variations and the seismogenic layer thickness (Fig. 11). The latter is typically estimated by using an isotherm that describes the depth up to which seismic events could originate. In previous studies isotherms between 573 K and 1073 K have been tested (Phillips, 1991; Knapmeyer et al., 2006; Plesa et al., 2018a). The 573 K isotherm marks the temperature at which quartz, the most ductile component of a granitic crust starts to show a plastic behavior (Schödl, 1998). Thus, this temperature is often associated with the bottom of the seismogenic layer on Earth. The 1073 K isotherm is more representative for a basaltic composition, as it marks the maximum depth of oceanic intraplate quakes (Bergman, 1986; Wiens and Stein, 1983). Since the majority of the martian crust is thought to have a basaltic composition, this value has been suggested to be more representative for determining the depth of the seismogenic layer on Mars.

Figure 11: Maps of the seismogenic layer at present day computed using the 573 K isotherm (panels a, b, and c) and the 1073 K isotherm (panels d, e, and f). Panels g, h, and i show histograms for the seismogenic layer thickness. Panels a, d, and g show the results obtained for the thin crust end-member case. Panels b, e, and h show the seismogenic layer thickness for the density dichotomy crust, and panels c, f, and i show the values that were obtained for the thick crust end-member. For orientation, the white contour lines show the 0 km level of surface topography.

In Fig. 11 the depth of the seismogenic layer is shown for the three thermal evolution models discussed in Section 3 that employ the crustal thickness variations illustrated in Fig. 1. The seismogenic layer was calculated by using the 573 K (Fig. 11a, b, and c) and 1073 K isotherms (Fig. 11d, e, and f), and the results are shown in Table 6. Similar to the models presented in Plesa et al. (2018a), due to their effects on the lithospheric temperatures, the variations of the crustal thickness control the seismogenic layer thickness variations. For the 573 K isotherm a clear dichotomy can be observed for the seismogenic layer thickness of the thin crust and thick crust end-members (Fig. 1a and b). For the density dichotomy crust, where variations in crustal density reduce the variations in crustal thickness, the seismogenic layer thickness is more homogeneous, but shows small values in Tharsis and Elysium provinces that are characterized by a thicker crust compared to the rest of the planet. For the 1073 K isotherm the crustal thickness dichotomy pattern is no longer visible for the thin crust model, as in this case this temperature is attained at a depth that is no longer sensitive to the temperature variations caused by the crustal thickness pattern. Due to the thicker crust and higher amount of crustal HPEs in the thick crust model, a dichotomy in the seismogenic
layer is still visible for the 1073 K isotherm. The crustal density dichotomy model shows a thin seismogenic layer in Tharsis and Elysium areas and otherwise a rather homogeneous distribution.

Table 6: Seismogenic layer thickness obtained the three models presented in Fig. 1 using a 573 K and a 1073 K isotherm. Min, median, and max show the minimum, median and maximum values attained in each model.

| Isotherm | thin crust | density dichotomy crust | thick crust |
|----------|-----------|-------------------------|------------|
|          | [km]      | [km]                    | [km]       |
| 573 K    | min       | 38.77                   | 28.71      | 33.74 |
|          | median    | 76.40                   | 75.83      | 58.52 |
|          | max       | 104.16                  | 104.16     | 109.19 |
| 1073 K   | min       | 143.42                  | 79.48      | 90.07 |
|          | median    | 219.15                  | 221.50     | 193.87 |
|          | max       | 265.29                  | 266.55     | 281.77 |

For the 573 K isotherm the seismogenic depth is much shallower compared to the 1073 K isotherm. The seismogenic layer thickness of the models presented in Fig. 11 show values between 29 and 109 km when the seismogenic layer is defined using the 573 K isotherm, while for the 1073 K isotherm the seismogenic layer thickness extends to depths of 282 km. Compared to the 573 K isotherm, the range of seismogenic layer thicknesses obtained with the 1073 K isotherm is nearly twice as large. As already shown by Plesa et al. (2018a) the range of seismogenic layer thickness increases for large crustal thickness variations. The values presented in Fig. 11 are smaller than the high density crust models (HC models) of Plesa et al. (2018a), but similar to the values obtained for the crustal thickness of Neumann et al. (2004) (NC models) and those for a density dichotomy crust (DC models). This is due to the fact that the HC models of Plesa et al. (2018a) have a thick crust (87.1 km on average) with a high amount of crustal HPEs and a cold mantle and lithosphere that lead to a thick seismogenic layer. Later, these models were excluded, as they were found to produce a much thinner elastic thickness than the south pole estimate due to the presence of a decoupling layer between the elastic cores of the mantle and crust (Plesa et al., 2018b). Furthermore, the average thickness of the HC models exceeds 72 km and is thus incompatible with the recent seismic data of the InSight mission (Knapmeyer-Endrun et al., 2021).

The seismogenic layer thickness is directly linked to the depth of seismic events, since deep seismic events could be indicative of a cold and thick lithosphere. For the three models shown in Fig. 11, some regions such as the Tharsis region, Elysium Planitia, and Arabia Terra region show a different seismogenic thickness depending on the exact model and the isotherm used. We note, however, that deep events are necessary to be able to distinguish between seismogenic depth distributions predicted from thermal evolution models (Plesa et al. 2018a). Shallow seismic events can occur in both thin and thick seismogenic layers. In general, deep seismic events would indicate that the 1073 K isotherm is more appropriate to define the seismogenic layer thickness. Specifically, deep seismic events in e.g., Arabia Terra could be indicative of a seismogenic thickness distribution such as the one observed for the density dichotomy crust (Fig. 11), while deep seismic events in Daedalia Planum around Arsia Mons and in the adjacent Terra Sirenum region could exclude the thick crust model. Deep seismic events in the northern part of the Tharsis province around Alba Mons would favor the thin crust model, while seismic events as deep as 280 km could only be obtained in Hellas basin in the thick crust model. However, source depths of marsquakes recorded by InSight lie at about 20–50 km below the surface (Brinkman et al., 2021; Stähler et al., 2021), but depth uncertainties remain large (Brinkman et al., 2021). Thus, currently none of the seismogenic layer distributions suggested by Plesa et al. (2018a) and shown in Fig. 11 can be excluded based on the depth of seismic events.

The seismogenic layer volume can be calculated from the seismogenic layer thickness and can be used to estimate an annual seismic moment knowing the strain rate from thermal evolution models.

\[ M_{\text{cum}} = \eta \dot{\varepsilon} V \mu \Delta t \]  

where \( \eta \) is the seismic efficiency with values between 0 and 1 that describe how much of the strain is released in form of seismic events compared to aseismic deformation. The strain rate \( \dot{\varepsilon} \) is estimated from thermal evolution models. \( V \) is the seismogenic layer volume, \( \mu \) is the shear modulus and \( \Delta t \) is the time interval used to compute the seismic moment.

Previous studies by Knapmeyer et al. (2006) and Phillips (1991) have used parametrized thermal evolution models and investigated the rate of planetary cooling to estimate an annual seismic moment. In a more recent study, Plesa et al. (2018a) used global 3D models and estimated the annual seismic moment distribution based on
the local contributions of strain rates associated with mantle cooling and convection. The contribution associated with convective stresses was found to be high in regions covered by a thick crust that leads to higher subsurface mantle temperatures and lower viscosities allowing for material to flow. The contribution associated with cooling stresses on the other hand was found to be high in area covered by a thin crust such as the northern hemisphere or large impact basins. In the absence of a thick insulating crust, these areas cool more efficiently and can produce higher cooling stresses compared to regions covered by a thick crust. While the seismic moment contributions from convective and cooling stresses are anti-correlated, given the fact that Mars is a stagnant lid planet and thus convective stresses are negligible in the shallow subsurface, the contribution from convective stresses is typically smaller compared to that of cooling stresses. Moreover, the contribution of convective stresses is entirely absent for the 573 K isotherm, as this isotherm would lead to a thin seismogenic layer.

The distribution of the annual seismic moment is shown in Fig. [12] For the 573 K isotherm the distribution of the annual seismic moment reflects the cooling pattern of the lithosphere that is controlled by the crustal thickness variations. Areas covered by a thick crust show a lower annual seismic moment budget, due to their slower cooling compared to areas covered by a thin crust. For the 1073 K isotherm, on the other hand, the annual seismic moment distribution is rather homogeneous. The contribution associated with convective stresses illustrates that a higher seismic moment can be attained in the southern hemisphere and the Tharsis and Elysium volcanic provinces, given their thicker crust that leads to warmer temperatures in those regions. The cooling stresses are more homogeneously distributed with slightly lower values in Tharsis and the southern hemisphere in particular for the density dichotomy and thick crust models (Fig. [12] and i, respectively). Nevertheless, when combining the contributions from cooling and convective stresses, the total contribution leads to a more homogeneous pattern, as it was also discussed by Plesa et al. (2018a).

The total available annual seismic moment budget can be used to compute a size-frequency distribution that often follows a Gutenberg-Richter law. The size-frequency distribution indicates the number of events that would be expected to occur over the course of a year with a seismic moment larger or equal to the largest assumed marsquake (seismic moment M\(\text{S} \geq 10^{26}\)Nm yr\(^{-1}\)). For the size-frequency distribution, the moment release obtained from global 3D thermal evolution models that include the recent constraints from the InSight data on core size and crustal thickness is shown in Fig. [13] Models indicate an annual cumulative seismic moment between \(5.19 \times 10^{16}\) and \(1.52 \times 10^{19}\) Nm, similar to previous values from thermal evolution calculations (Knapmeyer et al. 2006, Plesa et al. 2018a). Fig. [13] also includes previous estimates from lithospheric cooling computed using parametrized thermal evolution models (Knapmeyer et al. 2006, Phillips 1991), as well as from total slip on surface faults (Golombek et al. 1992, Golombek 2002). The seismic moment release of the Earth was obtained from Harvard Centroid Moment Tensor catalog between 1976 and 2013, while that of the Moon was derived from shallow moonquakes (Oberst 1987). The seismicity of Mars was derived based on InSight observations of marsquakes and extrapolated to the entire planet to account for uncertainties in detecting small and distant events (Banerdt et al. 2020).

The size-frequency diagram in Fig. [13] is sensitive to the largest seismic moment assumed. Values of the maximum possible seismic moment for Mars were estimated based on data from oceanic and continental intraplate quakes on Earth and lie around \(10^{20}\) Nm yr\(^{-1}\) (Phillips 1991, Golombek et al. 1992, Golombek 1994). Adopting this value and using the total cumulative moment release (Eq. (10)) the moment-frequency relation can be calculated. Additional uncertainties are related to the seismic efficiency \(\eta\) that lies between 0.025 and 1 for events on Earth (Ward 1998b), with values larger than 0.7 being representative for regions located at the border between North American and Pacific plates and small values indicating small strain regions, typical for central USA and northwest Europe (Ward 1998b). In addition, the cumulative moment release is proportional to the shear modulus, that for PREM varies between 26.6 GPa at the surface to 68.2 GPa at the base of the crust (Dziewonski and Anderson 1981).

It is important to note that the models of Plesa et al. (2018a) and those presented in Fig. [12] do not include the contribution of stresses produced by lithospheric flexure due to loading. Tensile-compressive stresses as well as shear stresses distribution in the martian lithosphere correlate with surface structures and can affect the seismic moment distribution in particular in areas such as Tharsis, Hellas Planitia, Argryre Planitia, Acidalia Planitia, Arcadia Planitia, and Valles Marineris. In addition, although the seismicity distribution Fig. [12] includes the contribution from convective stresses that reflect the presence of strong mantle plumes in the interior, it does not consider the contribution from magmatic processes that may be ongoing on Mars. This contribution may be specifically important in areas close to the large volcanic centers in Tharsis and Elysium, and may explain the observed seismicity in Cerberus Fossae. Thus future models that evaluate the present-day distribution of seismicity need to include these additional contributions.

Tectonic faults at the surface are important indicators of the internal stress distribution in the lithosphere (Banerdt et al. 1992, Carri 1974, Golombek and Phillips 2010, Wise et al. 1979). However, the distribution of
Figure 12: Spatial distribution of the annual seismic moment budget computed using the 573 K isotherm to define the seismogenic layer thickness (panels a, b, and c). Panels d, e, and f show the convective stresses contribution, panels g, h, and i the contribution associated with cooling stresses, and panels j, k, and l the total annual seismic moment budget computed using the 1073 K isotherm. Left column (panels a, d, g, and j) shows the results obtained for the thin crust end-member case, middle column (panels b, e, h, and k) shows the case with a density dichotomy crust, and right column (panels c, f, i, and l) shows the thick crust end-member case. The white and gray contour lines show the 0 km level of surface topography.
Figure 13: Comparison of the moment-frequency diagram for the Earth, Moon, and Mars. For Mars, the comparison includes the models presented in Fig. 1 (this study), previous seismicity estimates from Knapmeyer et al. (2006), Golombek (2002), Golombek et al. (1992), and Phillips (1991), as well as the values derived from the InSight data (Banerdt et al., 2020). Similar to Plesa et al. (2018a), the maximum seismic moment for a marsquake was assumed to be $10^{20}$ Nm and the slope was set to 0.625 (Knapmeyer et al., 2006), as suggested from the analysis of quakes occurring above the olivine-bridgmanite transition on Earth (Kagan, 2002).
present-day seismicity cannot be robustly estimated from the distribution of these features, as this would require knowledge of which faults are active today. The distribution of seismically active zones on present-day Mars could range from a nearly homogeneous one, if all faults are considered to be seismically active, to limited areas on the northern hemisphere and in Tharsis and Elysium, if only faults cutting Amazonian terrains are active today [Knapmeyer et al., 2006; Plesa et al., 2018a]. Thus, the localization of seismic events is essential to constrain the distribution of seismicity. Maps of the location of seismic events that were recorded by InSight could be used in future studies to discriminate between scenarios of seismicity distribution proposed by the analysis of various stress contributions and fault locations on the martian surface.

7 Conclusions and future work

The large amount of data and the diversity of data sets that are now available for Mars provide a unique opportunity to investigate the planet’s thermal evolution and constrain poorly known parameters such as the mantle viscosity, thermal variations in the interior, or the distribution of heat producing elements between the mantle and the crust. In particular, recent results of the InSight mission provide the most direct constraints for the martian interior. Crustal thickness values, the size of the core and information about the thickness of the thermal lithosphere can be combined with interior evolution models to constrain the thermal history and present-day state of the martian mantle and core.

Global geodynamic models show that the crustal thickness variations control the surface heat flow and elastic lithosphere thickness pattern. The present-day elastic lithosphere thickness at the north pole of Mars is one of the strongest constraints for thermal evolution models and indicates that the mantle contains less than 45% of the total heat production and/or that the polar cap has not yet reached elastic equilibrium.

Lithospheric temperatures show strong variations that correlate with the crustal thickness pattern. These variations lead to a seismic velocities pattern that can extend to depths of 400 km and deeper, depending on the exact crustal thickness model and crustal enrichment in HPEs. The seismic velocities variations due to temperature are larger than due to different compositional models for the mantle.

Thermal evolution models with a large core and a dry mantle viscosity can match the observed tidal deformation values. The size of the core indicates that the convection pattern in the mantle is characterized by several mantle plumes and downwellings, with stronger plumes preferentially focused in Tharsis due to a thick insulating crust at this location. Thus, the formation of the martian crustal dichotomy cannot be explained by an endogenous process that would require a low degree convection pattern, but is likely the result of the combined effects of large-scale impacts during the early martian history and subsequent interior dynamics.

The seismicity obtained from thermal evolution models that employ the latest crustal thickness and core radius estimates is compatible with the seismicity derived from InSight’s observations. The seismogenic layer thickness is sensitive to the crustal thickness variations and can be used to exclude thermal evolution models, if the depth of the events is known and exceeds the model predicted depth.

Several open questions, however, remain and require more modeling work and future observations. One of the largest unknowns that still remains is the surface heat flow of Mars. While estimates of the elastic lithosphere thickness at the north and south poles can be used to constrain the heat loss and the thermal state of the lithosphere at those locations, future models need to investigate scenarios, in which recent magmatic activity in Tharsis and Elysium is compatible with a large elastic thickness at the north pole of Mars.

Since the martian core is most likely liquid at present day, geodynamic thermal evolution models need to investigate whether a thermally driven dynamo can be reconciled with the duration of the martian magnetic field. While in a recent study by Greenwood et al. (2021), 1D thermal evolution models were found compatible with an internally generated magnetic field between 4.1 and 3.6 Gyr ago, it remains to be tested whether these scenarios are compatible with additional constraints on mantle cooling imposed by the large elastic lithosphere thickness at the north pole (Broquet et al., 2020) and by recent volcanic activity in Tharsis and Elysium (Neukum et al., 2004; Vaucher et al., 2009; Hauber et al., 2011).

Future models need to consider constraints from the Chandler Wobble, the movement of the pole away from the planet’s average rotation axis, that have been determined from radio tracking observations of Mars Odyssey, Mars Reconnaissance Orbiter, and Mars Global Surveyor (Konopliv et al., 2020). Since the Chandler Wobble period is sensitive to the rheology of the martian mantle, this would provide, in addition to the tidal quality factor Q, a valuable constraint for the mantle viscosity and thermal state of the interior.

Seismic velocities obtained from global geodynamic models need to be combined with seismic waves propagation to test the effect of variations in the mantle and lithosphere on estimated travel times. While in the convective mantle the variations in seismic velocities are typically small, temperature variations in the lithosphere may signif-
icantly affect the travel times of seismic waves, in particular for travel paths located on the northern and southern hemispheres. So far, the distribution of seismicity included only the contributions from cooling and convective stresses. Future models need to include stresses associated with topographic loads and evaluate the contribution from magmatic processes. Their predictions for the distribution of seismicity could be constrained with the observed locations of marsquakes.

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